Forebulge influence on deposition of the Cretaceous Castlegate Sandstone, Book Cliffs, Utah, U.S.A.

By

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ABSTRACT

Analysis of facies changes and paleocurrent directions of the incised valley fill of the Late Cretaceous Castlegate Sandstone supports a reinterpretation of the axis of the well-studied Castlegate Sandstone incised valley axis from an east-west trend to a north-south trend. Strata in the easternmost extent of the incised valley fill are reinterpreted to represent fill along the edges of the valley, rather than loss of fluvial energy at drainage terminations as previously interpreted. Drainage systems of the Castlegate Sandstone illustrate the influence of topography of the Sevier foreland basin on depositional patterns of the western extent of the Cretaceous Western Interior Seaway (KWIS) during lowstand events.

The Castlegate Sandstone was mapped from Tuscher Canyon, Green River, Utah to its easternmost expression to gather paleocurrent data and investigate morphologic changes within the incised valley fill. Early-stage valley fill, consisting of braided-stream facies, is concentrated in the western portion of the study area and paleocurrents in these deposits record flow to the south. Late-stage valley fill consists of meandering-stream deposits. The eastern portion of the incised valley fill is dominated by late-stage deposits with paleoflow to the west.

Detailed correlation of Castlegate Sandstone lowstand strata indicates forebulge exposure. Forebulge exposure decoupled the Sevier foredeep from the KWIS, allowing foreland basin topography to become the major influence on drainage systems during lowstand events. The north-south orientation of the axis of the Castlegate Sandstone incised valley reflects axial drainage of the foreland basin.

Reexamining lowstands of the Castlegate Sandstone, emphasizing facies reconstruction and paleoflow directions, illustrates the effect of the basin topography on the drainage patterns of

the KWIS. Recognizing that the incised valley expressed axial flow may lead to new interpretations about the stratal evolution of the foreland basin.

DEDICATION

For my parents, who taught me how to search for Ithaca.

"When you start on your journey to Ithaca,
pray that the road is long,
full of adventure, full of knowledge."
~C. P. Cavafy

ACKNOWLEDGEMENTS

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I am extremely grateful for the input and guidance from my advisor, Dr. Diane Kamola, on the development, implementation, and completion of this project. I would like to thank her for providing me with such an exciting project. I would also like to thank her for allowing me to control preparation and execution of the field study, which allowed me to become a more proficient scientist, even though it meant a second trip to the study area.

I would like to state my appreciation for my committee members, Dr. Anthony Walton and Dr. Doug Walker, who provided input and direction. I also would like to Dr. Diane Kamola, Justin Davis, and Benjamin Rickards for making the logistics of field work possible and enjoyable.

Without the support, guidance, and instruction of my officemates over the years I could not have accomplished this, so thank you. I appreciate your willingness to put up with my random questions and conversations, my ridiculous music, and my impromptu dance parties.

Finally, I want thank my family, friends, and Ben for their love, kindness, and support. I owe you all more than I can possibly put into words. Without you this would not have been possible.

TABLE OF CONTENTS

THESIS ARRANGEMENT

This thesis consists of two chapters and an appendix. Chapter 1 provides information on the geologic background of the area, reviews previous studies and subsequently named stratigraphy, and discusses research methods utilized in this study. Chapter 2 contains the data gathered during the study, interprets the data, and discusses the implications of the research. Chapter 2 is written in standard publication format to be submitted for publication to the American Association of Petroleum Geologists Bulletin. There will be some repetition of information presented in Chapter 1and the appendix with information presented in Chapter 2. The appendix is a compilation of field data presented as tables and cross sections.

CHAPTER 1

Introduction

The Late Cretaceous Mesaverde Group consists of an eastward-thinning clastic wedge that prograded into the Sevier Foreland Basin during and after the Sevier Orogeny. Two members of the Mesaverde Group, the Late Campanian Desert Member of the Blackhawk Formation and the Castlegate Sandstone of the Price River Formation, form progradational tongues of fluvial and marine strata (Fig. 1). These strata prograded basinward (eastward) and intertongue with the marine Mancos Shale to the east. The Desert Member and the Castlegate Sandstone crop out as an escarpment of Cretaceous strata known as the Book Cliffs of east-central Utah, USA (Fig. 2).

The Desert Member and Castlegate Sandstone east of Green River, Utah, are composed of several high-frequency sequences (on the order of 200,000 to 400,000 Ka; Van Wagoner, 1995). One sequence boundary exists in the Desert Member and at least three sequence boundaries are present in the Castlegate Sandstone (Van Wagoner, 1995). In addition to the Desert sequence boundary and three Castlegate sequence boundaries, Van Wagoner (1995) recognized a number of high-frequency sequence boundaries. These high-frequency sequence boundaries are interpreted to have formed either in response to minor eustatic changes (Van Wagoner, 1995, 1998) or to a relative sea-level fall linked to thrust activity (Yoshida et al., 1996, 1998).

Gill and Hail (1975) collected ammonites from several locations in the Desert Member, the Castlegate Sandstone, and the Buck Tongue of the Mancos Shale. Using ammonite dating provided by Fouch et al. (1983) and Obradovich (1993), Van Wagoner (1995) determined the age of the Desert sequence boundary to be 80 to 78 my. Deposition of the Castlegate Sandstone began around 79.7 Ma and lasted until 77.4 ± 0.5 Ma (Fouch et al., 1983; Gradstein et al., 1995; Aschoff and Steel, 2010). By dividing the age span by the number of sequences in the Desert Member and Castlegate Sandstone, Van Wagoner (1995) determined each sequence to last between 200,000 and 400,000 yr.

This study focuses on the outcrops of the Castlegate Sandstone east of Green River, Utah, to investigate the terminal patterns of the Castlegate Sandstone lowstand strata. The incised valley fill basally bounded by the Castlegate sequence boundary 2 is the primary focus of the study and was chosen because its preserved strata contains the most complete incised valley fill. Study of the drainage system of the lowstand systems tract (LST) deposits will emphasize how the morphology of the Sevier foreland basin impacted flow direction, and therefore sediment deposition, during lowstand events. Reconstructing such a framework requires accurate regional correlation and relies on careful analysis of paleocurrent directions. Field study of the incised valley fill within the Castlegate Sandstone requires detailed stratigraphic descriptions.

Geologic Background

The Sevier thrust belt developed during the Cretaceous Sevier Orogeny as a result of subduction of the Farallon plate beneath the western edge of the North American plate. Crustal loading and thrusting led to the formation of a retroarc foreland basin (Fig. 3) known as the Sevier foreland basin (Jordan, 1981; DeCelles and Giles, 1996). Foreland basins are elongate depressions in which the main depozone, the foredeep, is bound by the thrust belt and the forebulge (DeCelles and Giles, 1996).

The depth and width of the foredeep are directly related to the loading of the thrust sheet and the flexural rigidity of the lithosphere (Jordan, 1981). Strata fill of a foredeep forms an asymmetrical wedge that is thickest near the orogenic wedge and quickly thins towards the forebulge. Generally, the depozone of a foredeep is 2 to 8 km thick (DeCelles and Giles, 1996) and the Cretaceous Sevier foredeep was between 100 to 180 km wide (Currie, 2002; Liu et al., 2005). A forebulge is a potential zone of flexural uplift that forms in response to lithospheric rebound and creates a positive relief feature (Jacobi, 1981; Quinlan and Beaumont, 1984; DeCelles and Giles, 1996). Forebulge-derived sediment is transported into the foredeep and towards the craton (Crampton and Allen, 1995).

Sea-level rise beginning in the Aptian (Vail and Mitchum, 1979; Kauffman, 1984) caused the Circum-Boreal Sea to encroach from the north and the proto-Gulf of Mexico to transgress from the south. As a result, marine waters flooded the foreland basin and areas to the east forming the large epeiric sea known as the Cretaceous Western Interior Seaway (KWIS). The foreland basin was situated on the western edge of the KWIS and remained flooded until the Maastrichtian (Fig. 4). While it has been suggested the KWIS was as deep as 300 m (Kauffman, 1977; Jordan, 1981; White et al., 2002), water depth in modern submarine foredeeps (on continental crust) is commonly less than 200 m (DeCelles and Giles, 1996). Pang and Nummedal (1995) used lithofacies characterization to propose a water depth of less than 100 m on the western edge of the KWIS.

There is no widely accepted trigger for the relative fall in sea level recorded on the western edge of the KWIS. Two main schools of thought exist: tectonic effects that work independently of sea level (Schwans, 1995; Yoshida et al. 1996; Yoshida, 2000; Horton et al., 2004) and changes due to eustatic variations (Van Wagoner, 1995; Miller et al., 2004).

Tectonic events affect relative sea levels, generally on a long time scale. Erosion of the thrust belt is practically instantaneous with the uplift. When the thrusting stops, erosion becomes the main control on the lithosphere, leading to crustal relaxation and rebound associated with the denudation of the thrust belt. These thrusting cycles of loading, crustal relaxation, and subsequent rebound typically occur over 16.5 Ma periods (Schwans, 1995). While these events may be responsible for the large-scale sequence architecture of deposits within the foreland basin (Schwans, 1995; Yoshida et al. 1996), it is too long term to explain the sequences within the Castlegate Sandstone. However, Yoshida (2000) has related sequences of a frequency on the same order as the Castlegate to intraplate stress within one of the larger scale cycles. Provenance studies have found no link between provenance change and eustatic fall (Horton et al., 2004; Miall and Arush, 2001).

Tectonic influence plays a larger role closer to the orogen. The study area is located over ~160 km basinward of the thrust front, where the only tectonic influence occurs with initiation, emplacement, and cessation of a single thrust (Schwans, 1995). Duration of pulses of tectonics due to thrusting or cessation of a thrusting event is about one million years longer than relative changes in base level in foreland basins (Devlin et al., 1993). Climate and sea-level changes operate with frequencies of 10^3 to 10^5 years, while the tectonics influence longer term accumulation of strata over one to two million years (Van Wagoner, 1995). For this reason, Van Wagoner (1995) proposes climatic changes were responsible for the relative fall in sea level that led to deposition of the Castlegate Sandstone.

Because the rate of sea-level change was so quick (tens of meters per million years), glacioeustatic changes have also been cited as a possibility for a fall in relative sea level (Miller et al., 2004). Miller et al. (2004) suggest ephemeral ice sheets formed on Antarctica during the

Cretaceous, and these are responsible for sea level falls of ~25 m. This school of thought is not widely accepted.

The Sevier foreland basin and KWIS were subject to frequent eustatic changes. Thrusting events and subsequent erosion of the thrust belt resulted in constant shifting of the shoreline position along the western edge of the KWIS. Transgression of the sea occurred during periods of subsidence in response to crustal thickening due to thrusting (Kauffman, 1977; Jordan, 1981; Pang and Nummedal, 1995; Yoshida et al., 1996; Liu et al., 2005). Climate changes are probably responsible for high-frequency relative falls in sea level (Van Wagoner, 1995), but tectonic activity may influence longer-term regressions (Yoshida et al., 1998; Miall and Arush, 2001; Liu et al., 2005). The sea was too shallow to have more than a negligible effect on the flexure of the lithosphere (Jordan, 1981), but relative change in sea level played an important role in controlling sedimentation.

Four major thrusting events have been documented in the Sevier Orogeny between the Late Jurassic and the latest Cretaceous (DeCelles and Coogan, 2006). The Desert Member and the Castlegate Sandstone were deposited during the third episode of hinterland compression known as the Paxton thrusting event which lasted from 86 to 75 my (Currie, 2002; DeCelles and Coogan, 2006). Sandstones in the Blackhawk Formation and the Castlegate Sandstone are quartz-rich but contain detrital dolomite, feldspar, and clasts of chert and volcanic rocks (Lawton, 1986; Miall and Arush, 2001), denoting a source of low-rank meta-sedimentary, volcanic, and older sedimentary rocks (Hampson, 2010). This indicates the sediment source was Precambrian rock exposed in the Canyon Range thrust sheet (Van Wagoner, 1995; Robinson and Slingerland, 1998; DeCelles & Coogan, 2006). Petrographic, paleocurrent, and seismic data suggest a tectonically driven change in provenance occurs near the unconformity at the base of the Castlegate Sandstone (Yoshida et al., 1996; Miall & Arush, 2001) related to the unroofing of the Charleston-Nebo salient (Horton et al., 2004).

Stratigraphy

Lithostratigraphy

The Blackhawk Formation was first described as a series of progradational non-marine and fluvial to littoral deposits about 300 m thick (Young, 1955). The Blackhawk Formation is divided into six members (Fig. 1) representing progradationally to aggradationally stacked shoreface and deltaic sandstones that ultimately interfinger with the Mancos Shale to the east (Fouch et al., 1983). The youngest, the Desert Member, is composed of sandstone, shale, and coal-bearing rocks that thicken eastward from Green River to Thompson Canyon, where the unit begins to thin and pinch eastward until it interfingers with the Mancos Shale (Young, 1955).

The Castlegate Sandstone is an iconic, cliff-forming sandstone in the study of sequence stratigraphy that has been described in or is the subject of many stratigraphic studies (Young, 1955; Fouch et al., 1983; Van Wagoner, 1991; Olsen et al., 1995; Van Wagoner, 1995; Yoshida et al., 1996, 1998; Miall and Arush, 2001). At its type section in Price Canyon, the Castlegate Sandstone is a three-fold lithostratigraphic unit: the lower Castlegate Sandstone, the upper Castlegate Sandstone, and the Bluecastle Tongue (Fouche et al., 1983; Lawton, 1986). The Lower Castlegate is present in the study area, but the Upper Castlegate Sandstone and the Bluecastle Tongue are replaced laterally by the Buck Tongue of the Mancos Shale, Sego Sandstone, and Neslen Formation east of Green River, Utah (Lawton, 1983; Van Wagoner, 1995; Yoshida et al., 1996; Robinson and Slingerland, 1998; Miall and Arush, 2001). The Castlegate Sandstone is composed of high-energy fluvial to marginal-marine deposits.

In the eastern portion of the Book Cliffs the Castlegate Sandstone is overlain by marine deposits of the Buck Tongue, an interval of the Mancos Shale. A renewed increase in subsidence due to thrust activation caused a relative sea-level rise that flooded the Castlegate Sandstone and deposited marine shale of the Buck Tongue (Yoshida et al., 1996). The Buck Tongue thickens to the east until it transitions into the Mancos Shale as the Castlegate Sandstone pinches out (Young, 1955). While the Buck Tongue is present east of Green River, it is completely eroded to the west and north at the unconformity beneath the Sego Sandstone (Yoshida et al., 1996; Robinson and Slingerland, 1998).

Sequence Stratigraphy

Van Wagoner (1995) defined the sequence stratigraphy of the Desert Member and Castlegate Sandstone, partitioning the lithostratigraphic members into depositional units (Fig. 5). Van Wagoner (1995) determined that the Desert Member contained a complete sequence with at least one sequence boundary, a lowstand systems tract (LST), a transgressive systems tract (TST), and a highstand systems tract (HST). Van Wagoner (1995) proposed that the Castlegate Sandstone comprises a lowstand sequence set composed of at least three nested incised valleys. The elapsed time between sequences was roughly between 200,000 and 400,000 years (Van Wagoner, 1995).

North and west of the study area, the high-frequency sequences presented by Van Wagoner (1995) are correlated updip to the Castlegate Sandstone type-section near Price, Utah by McLaurin and Steel (2007). Yoshida et al. (1998) mapped the area west of and surrounding Green River, Utah to clarify the sequence stratigraphy and facies of the updip portion of the Desert Member and the Castlegate Sandstone.

Sea level maintained a slow rise during the majority of deposition of the Desert Member, but the LSTs of the Desert Member and the Castlegate Sandstone were deposited during a relative sea-level fall. Sequence boundaries in the Castlegate Sandstone can be traced from west of Tuscher Canyon to Horse Pastures. According to Van Wagoner (1995), the last fluvial expression of the lowstand system tract of the Castlegate Sandstone occurs near the Horse Pastures area. The down-dip extent of the lowstand shorelines, however, is placed farther east.

Previous Interpretations

The location of the Castlegate Sandstone lowstand shoreline, and its correlation with the fluvial portion of the Castlegate Sandstone lowstand, has been a source of geological debate (Van Wagoner, 1995; Yoshida et al., 1998; Robinson and Slingerland, 1998; Miall and Arush, 2001). The Castlegate Sandstone lowstand deposits are composed of eastward-thinning wedges of fluvial deposits that should transition into deltaic or marine deposits. No coeval sandstone shoreline deposits, however, have been described in literature.

Van Wagoner (1995) interpreted a belt of oolite deposits as the lowstand shoreline of the Castlegate Sandstone sequences and proposed that the fluvial expression of the LSTs never converged with the lowstand shoreline. Van Wagoner (1995) proposed that the semi-arid climate coupled with a short period of low sea level prevented a connection between lowstand rivers and the sea. He interprets the relative falls in sea level to have occurred so rapidly that river systems could not maintain a connection to the sea, inhibiting the supply of sediment necessary to produce sandy beach deposits. Instead, he posits the fluvial portion would thin eastward and transition into ephemeral swamps and lakes, and the LST shoreline would be muddy or oolitic (Van Wagoner, 1995). If the subsequent relative rise in sea level occurred rapidly enough, the fluvial system would remain detached from the LST shoreline.

Other studies propose that the Castlegate Sandstone drainage systems connected to the seaway, but that the shoreline deposits have been eroded (Yoshida et al., 1998; Miall and Arush, 2001). Yoshida et al. (1998) interpret the coeval lowstand shorelines of the Castlegate Sandstone to lie farther west than suggested by Van Wagoner (1995), and that the down-dip lowstand shoreline of the Castlegate Sandstone lies between Sagers Canyon and the Horse Pastures area (around Bull Canyon). Yoshida et al. (1998) argue that Castlegate Sandstone lowstand shorelines were removed as a result of beveling during the transgression of the Buck Tongue. They also propose that the shoreline deposits were eroded partially due to the emergence and migration of the forebulge that coincides with the thrusting event that led to the Buck Tongue transgression (Yoshida et al., 1996; Yoshida et al., 1998).

Van Wagoner (1998) counters this argument, identifying the preservation of rooting and dinosaur footprints as evidence that the non-marine portion of the Castlegate Sandstone was preserved, and indicating the Buck Tongue flooding surface was only slightly erosive or non-erosive during the Buck Tongue transgression. Miall and Arush (2001) maintain that a tectonic mechanism for the eastward thinning of the Castlegate Sandstone is preferred over Van Wagoner's (1995) sedimentologically based terminal fan model.

Facies analysis conducted in this study suggests the rivers maintained a connection to the sea. Generally, lowstand deposits show some minor tidal influence, including lags containing brackish-water to marine fauna, such as oysters, gastropods, and scaphopods. Other tidal indicators, such as double mud drapes, flaser bedding, wave ripples, and occasional sigmoidal cross-bedding are present in Castlegate Sandstone deposits. Channel sandstones often have basal lags that include logs with abundant *Teredolites*. *Teredolites* are borings from the shallow marine teredinid and pholadid bivalves (Bartsch, 1930). Logs riddled with *Teredolites* are often found in non-marine settings (Plint and Pickerill, 2007), but are present mainly in fluvial strata deposited during a period of transgression (Savarda et al., 1993). The presence of marine fauna and other tidal indicators throughout the fluvial portion of the Castlegate Sandstone suggests that drainage systems during lowstand events sustained a connection to the sea.

Previous studies have found that Castlegate Sandstone lowstand deposits have a general south-southeast paleoflow (Van De Graaff, 1972; Miall, 1993; Van Wagoner, 1995; Olsen et al., 1995). Erosive-based sandstones encased in marine shales that are associated with lowstand events, like the Sussex Sandstone and Shannon Sandstone (Tillman and Martinsen, 1984; Gustason et al., 1989), the Virgelle Member and Eagle Formation (Fitzsimmons, 1995), and the Rusty Member of the Ericson Sandstone (Martinsen et al., 1997) likewise show a general north-to-south paleoflow. These sand bodies also trend north-south. The southward trend of lowstand sand bodies throughout the Cretaceous North American foreland basin suggests that during lowstand events the topography of the foreland basin influences, and may control, drainage patterns on the western edge of the KWIS.

Forebulge Discussion

The position of the Sevier forebulge is difficult to locate because forebulges are erosive and migratory features. Forebulge location can be identified by the presence of unconformities (Jacobi, 1981; Quinlan and Beaumont, 1984) and regional onlap of strata onto the forebulge (Crampton and Allen, 1995; DeCelles and Giles, 1996; DeCelles and Currie, 1996). Pronounced forebulge uplift and migration occur with each new thrusting event (Stockmal et al., 1986), but the direction and rate of swell migration depend on a variety of factors, including the rheology of the lithosphere (Beaumont, 1978; Quinlan and Beaumont, 1984) and the elastic or visco-elastic response of the lithosphere (Jordan, 1981; Flemings and Jordan, 1990).

The location of the forebulge in the Sevier foreland basin during the Campanian is poorly constrained. As the Sevier thrust belt shifted eastward, the foreland basin and forebulge also migrated eastward. The Sevier foreland basin migrated roughly 250 km from east to west between the Late Jurassic and Early Cretaceous (Currie, 1997). Because of the difficulty of constraining all the factors necessary for predicting the forebulge position, several locations have been suggested (Currie, 1997; White et al., 2002; DeCelles and Coogan, 2006), but none is widely accepted.

Currie (1997) traced the forebulge from central Utah in the late Jurassic throughout the Cretaceous by palinspastically recreating the crest of the forebulge and using foreland basin isopach maps. Studies based on isopach maps of the strata fill of the foreland basin suggest the Campanian forebulge was located in central Colorado (Currie, 1997; White et al., 2002; DeCelles and Coogan, 2006). Some studies attempted to constrain the location of the Late Cretaceous forebulge using indicative facies (Yoshida, et al., 1998; Donaldson et al., 1999; Miall and Arush, 2001; Taylor et al., 2002). Localized ooidal ironstones up to 2 m thick have previously been recognized to be associated with forebulges and topographic highs (Plint et al., 1993; Taylor and Curtis, 1995; Donaldson et al., 1999; Taylor et al., 2002).Oolites interpreted by Van Wagoner (1995) as a lowstand shoreline of the Castlegate Sandstone are reinterpreted as the crest of the forebulge during deposition of the Castlegate Sandstone (Yoshida et al., 1998; Miall and Arush, 2001; Taylor et al., 2002).

Methods

Field study entailed detailed stratigraphic descriptions of 25 measured sections within 17 locations throughout the lowstand deposits of the Castlegate Sandstone. Vertical sections were measured to include all strata between the Desert sequence boundary and the base of the Buck Tongue. The base of the Buck Tongue, a regional flooding surface, was used as the stratigraphic datum. Sections measured between Tuscher Canyon and Sagers Canyon included strata below the Desert sequence boundary, interpreted as part of the Grassy HST. East of Sagers Canyon no Desert LST deposits have been documented. The HST of the Desert Member is indistinguishable from the Grassy HST, therefore only the uppermost parasequence of the Desert HST and the strata above were measured. Individual beds were described by grain size, lithology, bed thickness, sedimentary structures including cross-stratification, and fossil content (trace and body). Correlation of measured section was completed in the field. The stratigraphic framework and architectures presented in this study generally agree with those documented by Van Wagoner (1995).

The Book Cliffs provide continuous outcrops that are ideal for determining stratigraphic patterns as well as three-dimensional geometries of large- and small-scale sedimentary structures. The outcrops of the Castlegate Sandstone provide parallel- and perpendicular-to-flow views, critical to determining bar types and fluvial style (Bridge, 1993, 2006).

Paleocurrent data were obtained in the manner described by DeCelles et al. (1983). Axes of large- and small-scale trough crossbeds are generally aligned parallel to local primary flow direction, and paleocurrent data collected from such trough axes show a relatively small degree of scatter (DeCelles et al., 1983). Paleocurrents were taken from three-dimensional exposures of individual limbs of large-scale cross-strata from the basal parts of main channel fills. No paleocurrent data were acquired from the lowstand deposits of the Desert Member, as the emphasis of flow direction in this paper is focused on the Castlegate Sandstone lowstand drainage systems.

Figures

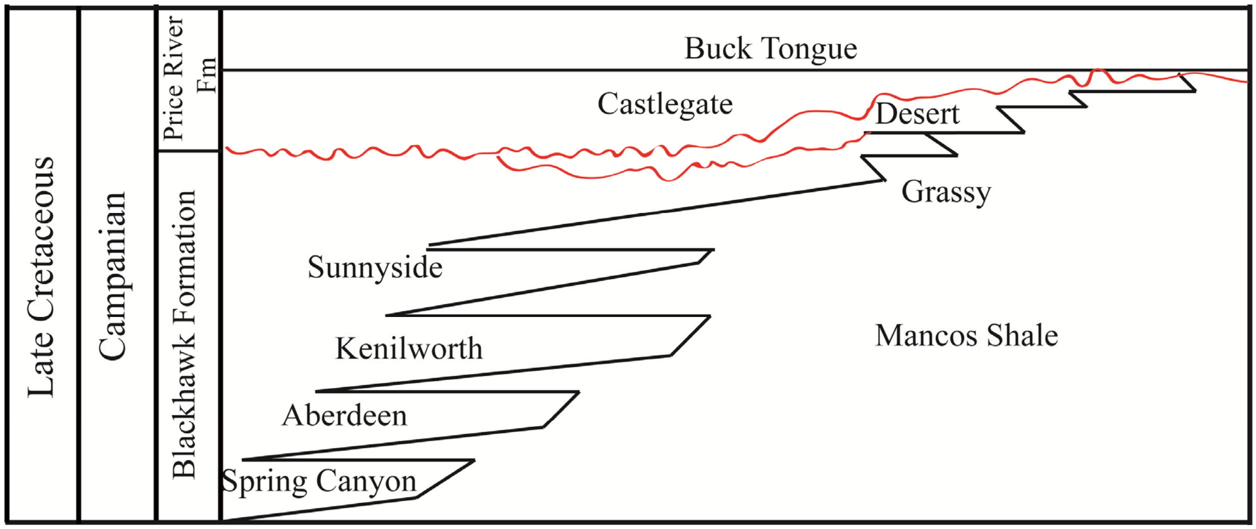

Figure 1: Campanian strata in eastern Utah include the Blackhawk Formation and the Castlegate Sandstone. These units are cut by sequence boundaries (red lines). The figure has been modified from Van Wagoner (1995).

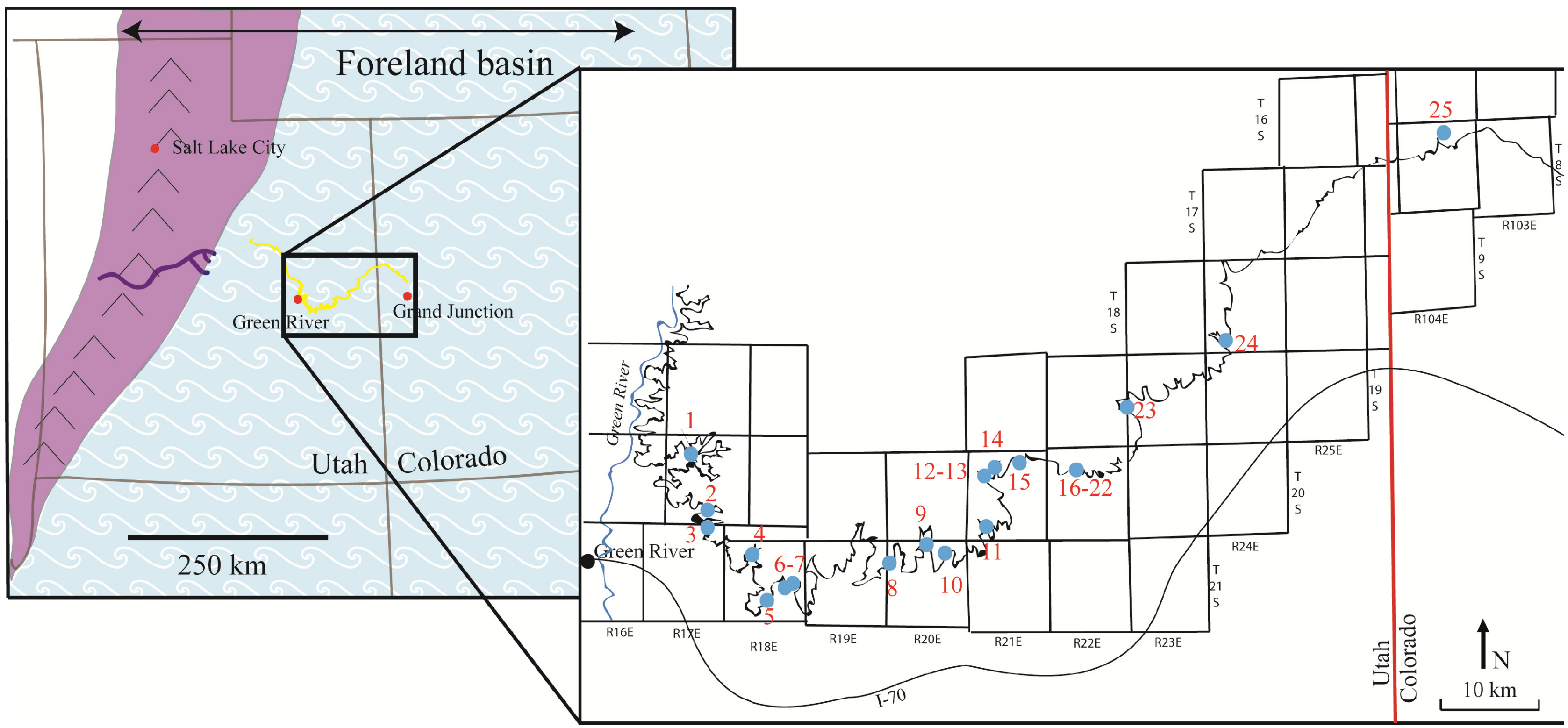

Figure 2: The study area is part of the Southern Book Cliffs that extends from Tuscher Canyon, Utah to West Salt Creek, Colorado. The outline represents the outcrop of the Castlegate Sandstone.

1: Tuscher Canyon
2: Coal Canyon
3: Stubb Canyon
4: Horse Canyon
5: Hatch Mesa
6: Floy Canyon
7: Floy Canyon North
8: Blaze Canyon
9: Thompson Canyon
10: Jeep Trail
11: Sagers Canyon
12: Sagers Canyon East
13: Bull Canyon A
14: Bull Canyon B
15: Strychnine Wash
16: Horse Pastures 1
17: Horse Pastures 2
18: Horse Pastures 3
19: Horse Pastures 4
20: Horse Pastures 5
21: Horse Pastures 6
22: Horse Pastures 7
23: Big Hole Wash
24: Sulpher Canyon
25: West Salt Creek Wash

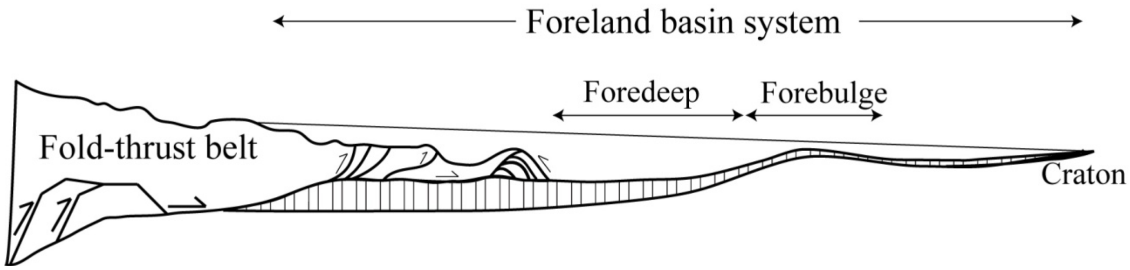

Figure 3: Retro-arc foreland basins have a foredeep that bound by the fold-trust belt and the forebulge. The figure is modified from DeCelles and Giles (1996).

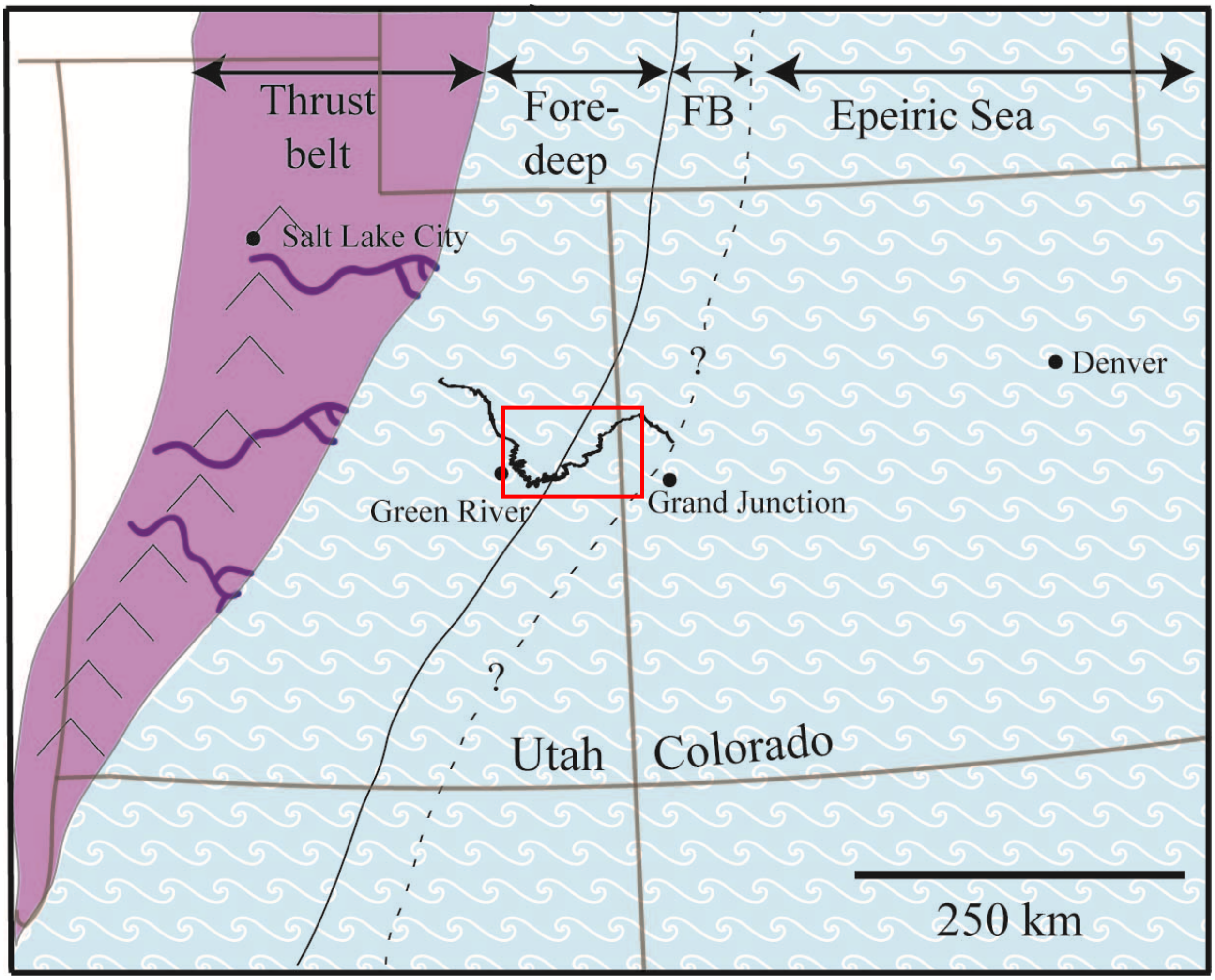

Figure 4: The Sevier foreland basin was situated on the western edge of the KWIS. The dashed line represents the unknown western edge of the forebulge. The red box denotes the study area. FB: forebulge.

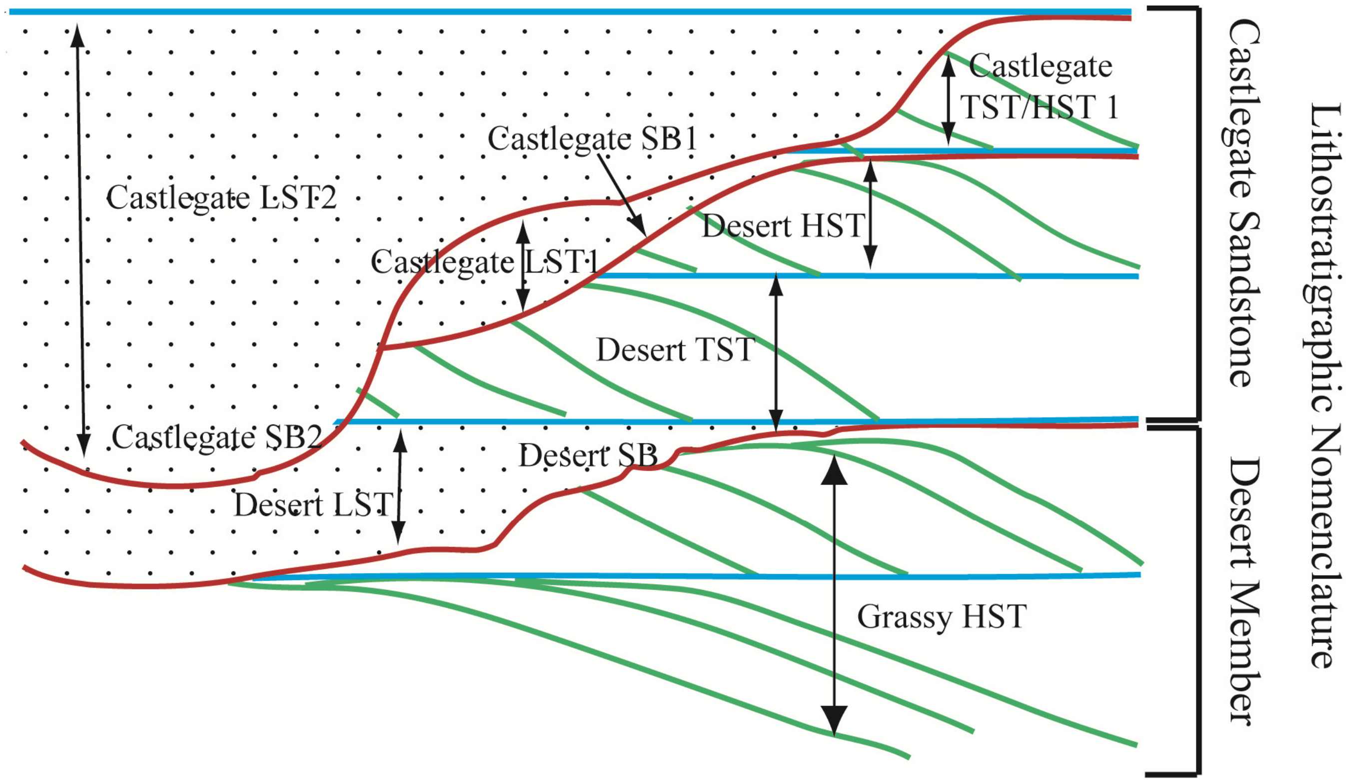

Figure 5: The wheeler diagram of the Desert Member and the Castlegate Sandstone is used to compare lithostratigraphic interpretation and sequence stratigraphic interpretation. Lithostratigraphic names are on the right side of the figure. Red lines represent sequence boundaries (SB), blue lines represent parasequence boundaries, and green lines represent normal marine progradation.

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CHAPTER 2

Introduction

The Sevier foreland basin played a significant role in sedimentation on the western margin of the Cretaceous Western Interior Seaway (KWIS). Drainage systems flowed east from the Sevier thrust belt towards the shoreline of the KWIS (DeCelles and Coogan, 2006). During lowstand events major drainage systems incised and extended eastward into the foreland basin. The Castlegate Sandstone lowstand strata contains deposits of fluvial systems that flowed east from the thrust belt, but also contain deposits that record drainage to the south-southeast (Van Wagoner, 1995).

The Castlegate Sandstone is part of the Upper Cretaceous Mesaverde Group (Fig. 1), an eastward-thinning wedge of strata deposited in the foredeep of the Sevier foreland basin (DeCelles and Currie, 1996). The majority of the Castlegate lowstand record is found in eastern Utah and consists of braided-stream deposits that transition into meandering-stream deposits to the east. The lowstand deposits are interpreted as fill of an incised valley (Van Wagoner, 1995). No coeval lowstand shoreline deposits, however, have been recognized basinward of the valley fill, either in outcrop or subsurface. Van Wagoner (1995) interprets the lack of correlative shoreline deposits as evidence that the drainage systems terminated before reaching the shoreline. The interpretation is widely contested (Nummedal et al., 1995; Yoshida et al., 1998; Miall and Arush, 2001; and others).

This study focuses on exposures of the Castlegate Sandstone between Tuscher Canyon, near Green River, Utah, and West Salt Creek, near Mack, Colorado, to explore lowstand drainage patterns within the incised valleys. Castlegate lowstand strata were selected for the study due to the presence of excellent outcrops in the southern Book Cliffs and their location within the foreland basin and near the western edge of the KWIS. In these locations, the Castlegate lowstand fluvial deposits exhibit a south-southeast paleoflow (Van Wagoner, 1995) even though proximal to the sediment source, the thrust belt, Castlegate Sandstone rivers flowed east (Van de Graaff, 1972; Lawton, 1983; Olsen et al., 1995). This study employs paleocurrent data and facies analysis to test the various hypotheses regarding the eastward thinning of the Castlegate Sandstone east of Green River, Utah.

This paper proposes that a relative fall in sea level restricted the influence of the KWIS on sediment dispersal and flow direction in the Castlegate lowstand strata. Following the relative fall in sea level, the topography of the foreland basin became a dominant influence on the drainage systems within the eastern extent of Castlegate Sandstone deposition.

Recognizing the impact of the topography of the foreland basin on drainage patterns in the Castlegate Sandstone lowstand strata improves our understanding of the complexity of the interplay between the KWIS and the foreland basin, especially during lowstand events. Paleogeographic reconstructions of the KWIS will differ in highstand and lowstand events. Understanding these differences will help predict the location of lowstand channel and deltaic deposits in underfilled basins.

Background Geology

The Sevier thrust belt developed during the Cretaceous Sevier Orogeny as a result of subduction of the Farallon Plate and the subsequent development of a magmatic arc along the western edge of the North American plate. Thrusting led to crustal loading and to the formation of a retroarc foreland basin known as the Cretaceous or Sevier foreland basin (Jordan, 1981; DeCelles and Giles, 1996). Foreland basins are elongate depressions that form on the lithosphere as a result of loading of a thrust belt. The lithosphere reacts to loading in a manner similar to how loading on an elastic beam overlying a fluid reacts to loading (Turcotte and Schubert, 1982). A major downwarp, the foredeep, forms adjacent to the load and is the major depozone of the foreland basin (DeCelles and Giles, 1996). A minor zone of potential flexural uplift, the forebulge, forms in isostatic response to the load and flanks the side of the foredeep that is not adjacent to the thrust belt (Turcotte and Schubert, 1982). The foreland basin on the western coast of North America (Fig. 2) acted as a topographic low which was inundated during transgression of the Circum-Boreal Sea from the north and the proto-Gulf of Mexico from the south (Jordan, 1981; Kauffman, 1984). Loading and subsequent erosion of the thrust belt influenced the subsidence and sedimentation in the western part of the KWIS (Jordan, 1981).Tuscher Canyon, the westernmost canyon analyzed in this study, is located approximately 180 km (110 mi) from the eastern edge of the Santaquin culmination, a large antiformal duplex of the thrust belt, a potential sediment source for the Castlegate Sandstone (Horton et al., 2004) (Fig. 3).

Deposition of sediment shed from the thrust belt during the Middle to Late Cretaceous partially occurred within the foredeep of the foreland basin (DeCelles and Currie, 1996). Foredeeps are preserved as an asymmetrical clastic wedge that thickens toward the orogenic wedge and thins toward the forebulge. As described by DeCelles and Giles (1996), the foredeep is the depression between the thrust belt and the forebulge. The Cretaceous Sevier foredeep was between 100 to180 km (60 to110 mi) wide (Currie, 2002; Liu et al., 2005).

The study area for the Castlegate Sandstone is restricted to the eastern part of the Campanian Sevier foredeep. As used in this paper, “foredeep” refers to the eastern portion of the Campanian Sevier foredeep, within 50 km (30 mi) of the forebulge. The forebulge separates the foredeep from the craton (Jacobi, 1981; Quinlan and Beaumont, 1984; DeCelles and Giles, 1996), and sediment derived from an exposed forebulge is transported both into the foredeep and towards the craton (Crampton and Allen, 1995).

The position of the forebulge is difficult to determine because forebulges are migratory features. Because of the difficulty of constraining all the factors necessary for predicting the forebulge position during deposition of the Castlegate Sandstone, several locations have been suggested but none is widely accepted. The Late Cretaceous forebulge has been placed in central Colorado (DeCelles and Giles, 1996; White et al., 2002; DeCelles and Coogan, 2006) based on palinspastic reconstructions of isopach maps of sediment thickness. In contrast, facies-based studies propose the location of the Campanian forebulge to be farther west, near the Colorado-Utah border (Yoshida, et al., 1998; Miall and Arush, 2001; Taylor et al., 2002). If the Campanian forebulge was located near the present day Colorado-Utah border, Castlegate Sandstone lowstand deposition would have onlapped the western flank of the forebulge (Fig. 4).

Stratigraphy

The Castlegate Sandstone is part of the Campanian Mesaverde Group, an eastward-thinning siliciclastic wedge sourced from the Sevier orogenic belt and deposited in the Sevier foreland basin. The Mesaverde Group contains the Blackhawk Formation, the Castlegate Sandstone, and the Buck Tongue of the Mancos Shale. A regional disconformity separates the Castlegate Sandstone from the underlying Desert Member of the Blackhawk Formation. A regional transgressive surface separates the Castlegate Sandstone from the base of the Buck Tongue of the Mancos Shale (Spieker, 1949) (see Fig. 1).

The Castlegate Sandstone has been described in several stratigraphic studies (Young, 1955; Fouch et al., 1983; Van Wagoner et al., 1990; Olsen et al., 1995; Van Wagoner 1995; Yoshida et al., 1996; Miall and Arush 2001). Gill and Hail (1975) collected ammonites from several locations within the Castlegate Sandstone and the Buck Tongue of the Mancos Shale. Using Gill and Hail's (1975) ammonite data and ammonite data provided by Fouch et al. (1983) and Obradovich (1993), Van Wagoner (1995) estimated the age of the lowstand strata of the Castlegate Sandstone to be between 79 to 77 my.

At its type section in Price Canyon, the Castlegate Sandstone is a three-fold lithostratigraphic unit: the lower Castlegate Sandstone, the upper Castlegate Sandstone, and the Bluecastle Tongue (Fouch et al., 1983; Lawton 1986). East of Price Canyon, the Upper Castlegate Sandstone, and Bluecastle Tongue are replaced laterally by the Buck Tongue of the Mancos Shale, the Sego Sandstone, and the Neslen Formation (Lawton, 1983; Van Wagoner, 1995; Yoshida et al., 1996; Robinson and Slingerland, 1998; Miall and Arush, 2001). A disconformity transects the Castlegate Sandstone, separating deposits within the study area physically and temporally from the Castlegate Sandstone at its type section (Van Wagoner, 1995).The Castlegate Sandstone within the study area consists of a sheet-like sandstone formed by an amalgamation of fluvial deposits that lacks any internal zones of continuous overbank deposits. The sandstone is interpreted to represent a braided-stream complex that was tens of kilometers in width and extended down dip for hundreds of kilometers (Van Wagoner, 1995).

Deposition of the Buck Tongue coincided with a major marine transgression (Van Wagoner, 1995; Yoshida et al., 1996). The Buck Tongue is expressed as an eastward-thickening tongue of marine shale that transitions into the Mancos Shale (Young, 1955). Although the Buck Tongue is present to the east of Green River, it is completely eroded to the west and north by the unconformity that underlies the Sego Sandstone (Van Wagoner et al., 1990; Yoshida et al., 1996; Robinson and Slingerland, 1998).

Sequence Stratigraphy

The sequence stratigraphy of the Castlegate Sandstone, east of Tuscher Canyon, is complex and described in detail by Van Wagoner (1995) (Fig. 5). The regional disconformity underlying the Castlegate Sandstone can be correlated over 100 km (60 mi) north into the Wasatch Plateau (Spieker, 1949; Young, 1955; Van Wagoner, 1995). This disconformity is interpreted as a sequence boundary (Van Wagoner et al., 1990, 1995; Olsen et al., 1995; Yoshida et al., 1996). Where the disconformity is not erosive, it is an interfluve surface that sometimes is distinguished by subaerial exposure represented by a *Glossifungites* or *Scoyenia* ichnofacies that sits on the top of the underlying lower shoreface deposits. Subaerial exposure surfaces in the Castlegate Sandstone may also display rooting and dinosaur footprints (Van Wagoner, 1995, 1998).

In the study area, the basal Castlegate Sandstone sequence boundary is overlain by complex stratal successions containing nested incised valley fill sandstones. The sheet-like sandstones are up to 80 km (50 mi) from east to west and make up a complex succession of lowstand deposits containing braided stream deposits and marginal marine deposits (Van Wagoner, 1995). Van Wagoner (1995) interprets these lowstand deposits to be a lowstand sequence set. Deposits from the lowermost incised valley system, underlain by sequence boundary 1, are only locally present and have been mostly removed and covered by a second valley system, which is present throughout the study area (Van Wagoner, 1995). The second valley system, underlain by sequence boundary 2, contains the best paleocurrent directions and valley-fill exposures.

The nested incised valleys form a wedge of eastward-thinning valley fill (Olsen et al., 1995; Van Wagoner, 1995; Yoshida et al., 1998). The Castlegate Sandstone lowstand strata are up to 35 m (115 ft) thick in the western portion of the study area, but thin to less than1 m (3 ft) at the most eastern expression. Fluvial strata in the west are dominated by high-energy deposits, while the strata in the east reflect lower energy depositional systems (Van Wagoner, 1995).

Regional correlation and paleogeographic maps show an eastward trend to the Castlegate Sandstone drainage systems (Van Wagoner, 1995). The drainage systems are shown to terminate before they reach the lowstand shoreline. Time-equivalent shallow marine strata are not associated with the terminal ends of the Castlegate Sandstone lowstand strata. Van Wagoner (1995) interpreted the Castlegate Sandstone lowstand strata to be the fluvial portion of a terminal alluvial megafan system in which braided rivers transitioned into ephemeral lakes and swamps in a semi-humid to arid climate. Substantial well log data as far as 200 km (125 mi) east of the Colorado border show no evidence of a sandstone lowstand shoreline; Van Wagoner (1995) concluded that the lowstand shoreline was either preserved as a carbonate oolite belt or a mudstone-on-mudstone boundary (Van Wagoner, 1995). Outcrops of ferrous oolites are traced as the eastern-most expression of the Castlegate Sandstone lowstand strata. Van Wagoner (1995) projected that the ooidal ironstone outcrops as a thin belt that paralleled the paleoshoreline and interpreted the ooidal ironstone belt as a detached Castlegate shoreline.

Van Wagoner's (1995) interpretation of the terminal patterns of the Castlegate Sandstone lowstand strata has been challenged frequently (Yoshida et al., 1998; Miall and Arush, 2001; Adams and Bhattacharya, 2005). Yoshida et al. (1998) propose lowstand shorelines were deposited within the eastern portion of the study area but were removed by beveling during the transgression associated with the Buck Tongue. Previous studies that generally agree with Van

Wagoner's (1995) sequence stratigraphic framework often suggest a tectonic mechanism, rather than sedimentary processes, to explain the distal thinning of the Castlegate Sandstone. Yoshida et al. (1998) propose that outcrops of lowstand strata of the Castlegate Sandstone do not parallel the trend of the channel fill sandstone. They interpret the apparent downdip decrease in the erosional relief of the Castlegate Sandstone to be an oblique transect across parallel, unrelated channel-fill sandstones of varying size (Yoshida et al., 1998) instead of a single fluvial system as previously proposed. Miall and Arush (2001) argue that the eastern fluvial portion of the Castlegate Sandstone was erosionally removed by uplift caused by events like the uplift of the Uncompahgre Uplift or incipient Laramide movement.

Others studies disagree with Van Wagoner's (1995) sequence stratigraphic correlations, maintaining the Castlegate Sandstone east of Tuscher Canyon is not part of a lowstand sequence set (Nummedal et al. 1992; Pattison, 1994; Adams and Bhattacharya, 2005; Pattison, 2010). These alternative interpretations correlate the Castlegate Sandstone fluvial deposits with shoreface deposits in the underlying Desert Member. Such an interpretation places the strata in a different stratigraphic framework and proposes that deposition occurred during the falling stage systems tract (Nummedal et al., 1992; Pattison, 1994; Adams and Bhattacharya, 2005; Pattison, 2010).

Castlegate sequence boundaries mapped in this study generally follow Van Wagoner's (1995) with minor deviations. Van Wagoner (1995) maintains three sequence boundaries are present within the Castlegate lowstand sequence set, all of which appear to correlate to the Buck Tongue flooding surface. This study proposes that only Castlegate sequence boundary 2 correlates to the Buck Tongue flooding surface. Only two sequence boundaries in the Castlegate Sandstone were mapped in this study, but the sequence boundaries are correlated to separate flooding surfaces. Documented hummocky cross-stratified beds sit between Castlegate sequence boundary 1 and Castlegate sequence boundary 2 (Van Wagoner, 1995). The presence of hummocky cross-stratification between the sequence boundaries indicates the occurrence of a transgression between deposition of Castlegate LST 1 and Castlegate LST 2 (Fig. 6).

Methods

The incised valley fill basally bounded by Castlegate sequence boundary 2 is the primary focus of the study. This fill is the most extensive and best preserved of the Castlegate Sandstone lowstand strata (see Fig. 5). Reconstructing the framework of the terminal patterns of the Castlegate Sandstone requires accurate, regional correlation and careful analysis of paleocurrent directions.

Field study entailed detailed stratigraphic descriptions of 25 measured sections within 17 locations throughout the lowstand deposits of the Castlegate Sandstone east of Tuscher Canyon. The Fluvial facies within the incised valley fill of the Castlegate Sandstone were described and analyzed to distinguish early incised valley fill from late incised valley fill. Paleocurrent directions were analyzed to investigate the overall flow patterns of the Castlegate Sandstone. The facies interpretations presented here are generally consistent with the facies described by Van Wagoner (1995).

Data

Facies:

Facies one consists of upward fining, fine- to medium-grained sublithic arenite. Stratification is dominated by large scale tabular to wedge cross-bedding (Fig. 7) and the cross-beds range from 35 cm to 1 m in thickness and form bedsets up to 8 m thick. Soft-sediment deformation may be present at the tops of beds. Within the bedsets, beds display upward fining and an upward decrease in bed thickness. The bedsets are bounded by erosional surfaces measuring up to 1 m of relief. Deposits above the basal erosional surface may contain a lag of pebbles, rip-up clasts, wood casts with *Teredolites*, and less commonly, bone and macerated plant fragments. When preserved in its entirety, current-ripple cross lamina may be preserved at the top of the bedset where preserved in its entirety. Wave-ripple laminae occur randomly. Cross laminae may be interbedded with silt interbeds or lenses 2 to 4 cm thick that may be disturbed by horizontal burrows up to 2 mm in diameter. No tidal influence is recorded in the sandstone.

Large-scale tabular cross-bedding indicates migration of two-dimensional megaripples up to 1 m in height. Large-scale cross-bedding along with upward fining, the decrease in bedform size within each bedset, and the lack of mud and lower energy indicators, are consistent with deposition of bars in a high energy, but not highly sinuous, river. The bedsets are composed of large, sigmoidal- to wedge-shaped bars (Fig. 8). This facies is interpreted as braided-river deposits. These deposits generally occur in the lowest10 m of strata above the Castlegate sequence boundary 2 in the western portion of the study area. Facies one deposits are not present in the eastern portion of the study area

Facies two consists of thin- to medium-bedded, fine-grained sublithic arenite. The grains are well rounded. Bedsets are up to 3 m thick, and are bound by erosional surfaces. Most bedsets have a basal lag of rip-up clasts and may contain wood casts with *Teredolites*. Rip-up clasts measuring up to several centimeters in diameter consist of sandstone and siltstone. Finer grained rip-up clasts maybe laminated. Wedge and compound cross-strata up to 40 cm thick occur at the base of each bedset. Beds pinch-out in the downstream direction. Beds thin towards the top of a bedset and may contain flaser-bedded current and wave ripples. Mud drapes and reactivation surfaces are common throughout the strata. Locally, *Skolithos* and *Planolites* are present in bedsets in the upper few meters of the incised valley fill. Van Wagoner (1995) identified *Arenicolites* in this facies.

Through-going, uniformly dipping surfaces that extend from the top to the base of some bedsets are associated with beds of this facies. The surfaces dip in the same direction and cross multiple sections of beds. These inclined surfaces are interpreted as lateral accretion surfaces (Fig. 9). The lateral accretion surfaces may be traced laterally to where the dip of the surface changes direction, showing the meander of a channel.

Skolithos, *Planolites*, *Arenicolites*, and *Teredolites* have been associated with brackish water deposits (Gingras *et al*. 1999). The upward-fining grain-size, upward thinning of bed thickness, abundance of lateral accretion surfaces, along with the presence of brackish water trace fossils, indicate these sandstones were deposited as fill of meandering channels. Flaser bedding, mud drapes, wave ripples, and brackish water trace fossils are evidence of tidal influence in the meandering stream deposits.

Meandering stream deposits are associated with gray, brown, and black siltstones interbedded with thin sandstones, facies three. The siltstones are poorly bedded and range from 5 cm to more than 1 m in thickness. Sandstone beds range from 2 cm to 30 cm in thickness. Siltstone beds become more abundant and sandstone beds become less abundant up section in rocks of this facies. Disseminated organic matter less than 1 cm in diameter occurs randomly in the sandstones and siltstones. The tops of sandstone beds may display roots up to a millimeter thick. Many siltstone beds are truncated and overlain by thin beds of current rippled and bioturbated sandstone. Ripple-cross stratification is rarely preserved within the siltstone.

Depending on the quality and extent of the outcrop, siltstones may be traced laterally to correlative channelized deposits. The interbedded sandstone and siltstone are interpreted as overbank and floodplain deposits. This facies represents crevasse splays and levee deposits that can be traced laterally to, and are associated with, channel belts.

Facies four consists of thick siltstone beds up to 4 m thick. The facies is rare but present near the top of the valley fill in the western portion of the study area. These deposits are either massively bedded or mottled and contain abundant organic matter, root churning, and blocky ped structures that transition into crumbly ped structures. Preserved roots generally measure between 2 and 5cm, but may be up to 15cm long (Fig. 10). A distinct vertical trend in color is recognized from dark gray at the base to light gray at the top. A gray-white layer only a few centimeters thick may be present near the top of the section. These deposits are interpreted to be paleosols.

The easternmost outcrops of the Castlegate Sandstone lowstand strata contain facies (facies five and facies six) that are only present about 30 km (18 mi) west of Grand Junction. Facies five consists of 30 cm of thinly bedded silty sandstone containing abundant organic matter and plant fossils up to 1 cm in diameter, including *Equisetales* (Fig. 11). Large vascular plant fragments are fragile and susceptible to disarticulation during transport, and the fragments of preserved plant matter in this area indicate in situ growth (Van Wagoner, 1995). The silty sandstone beds are discontinuous. Thin sections show palynomoprhs of fresh water to brackish-water vascular plant fragments and Pteridophyte spores (Van Wagoner, 1995). In situ fresh water to brackish-water plant fragments and organic matter were deposited in a fresh to brackish water marsh environment.

Facies six consists of cross-bedded ooidal ironstones and overlies facies five (Fig. 12). As with facies five, facies six is limited in its extent and is restricted to the easternmost part of the study area. The facies was measured at West Salt Creek to be up to 2 m thick. Facies six contains vague bedding and consists of ooids and quartz grains in a dark red matrix. Thin section analysis by Van Wagoner (1995) and Taylor et al. (2002) shows that the nuclei of the ooids consist of quartz, ammonite, or clay minerals and that the laminae are composed of apatite or iron oxides, such as berthierine. Most ooids have been partially to wholly dolomitized (Taylor et al., 2002). The ooids formed in high energy, shallow marine environments, and are interpreted to have formed in shoals that were exposed to periodic storm influence (Taylor et al., 2002) and as a detached lowstand shoreline (Van Wagoner, 1995).

Paleocurrent analysis:

Paleocurrent measurements were taken in the manner described by DeCelles et al. (1983), preferentially from beds displaying three dimensional, large-scale tabular cross-bedding. Cross bedding in the upper 5 m of the Castlegate Sandstone is typically bioturbated and weathered, providing fewer paleocurrent data. The overall trend in paleocurrent direction and number of paleocurrents gathered for each canyon are presented in Table 1. Data are plotted in 10° rose diagrams, and the vector mean was calculated for each canyon (Fig. 13). The magnitude of the vector mean is representative of the data scatter: small data scatter generates a vector mean with a large magnitude.

Between Tuscher Canyon and Sagers Canyon the cross-beds display a southerly trend in paleocurrent data, which ranges between 190^{o} and 130^{o}with an overall trend of 162^{o}. The magnitude of the trend line reflects a low dispersion of data. Measurements taken from the top few meters of deposits in the western portion of the study area maintain a general south-southeast paleoflow, but display greater paleocurrent scatter.

Between Sagers Canyon and the Horse Pastures area (the easternmost fluvial expression of the Castlegate Sandstone) the general paleoflow shows a westward trend of 267°. Most paleocurrent directions gathered in this portion of the study area trend west between 180° and 0°, but some paleocurrents record easterly flow. The magnitude of the vector mean of paleocurrents is small and the data scatter is high in individual canyons because paleocurrent directions can vary up to 180° per location. Compilation of all paleocurrent data in these canyons shows an overall trend line with an average magnitude which trends at 268°. Deposits with a consistent southern trend in paleoflow are associated mostly with the braided stream facies, while the meandering systems display scatter of higher paleocurrent data.

Interpretation

Braided-river deposits with no tidal influence are common in the western portion of the incised valley (between Tuscher Canyon and Blaze Canyon). The deposits stack vertically in a strongly aggradational pattern to form multistory sandstones (Fig. 14). The small dispersion of paleocurrent data indicates the braided rivers were a low-sinuosity system, with straight to slightly sinuous channels.

Braided-river deposits are overlain by meandering-stream deposits in the western part of the study area. Meandering stream deposits make up the entire eastern portion of the incised valley fill. The increase in preserved cross-cutting lateral-accretion surfaces in the upper portion of the incised valley fill illustrates that fluvial systems became increasingly sinuous as the system aggraded. Meandering stream deposits stack to form single- and multi-story architecture. Towards the top of the incised valley fill tidal influence becomes common, the thickness of individual meandering stream deposits decreases, and lateral accretion becomes increasingly common.

In Horse Pastures, the easternmost expression of the Castlegate Sandstone lowstand deposits, channelfills are only a few meters thick while overbank and floodplain deposits measure up to0.5 m thick. Throughout the incised valley fill, floodplain and overbank strata are progressively more abundant up section, and siltstones begin to show signs of paleosol formation, transitioning from blocky pedosols to crumbly pedosols. This reflects an increase in the frequency of flooding events up section.

Based on paleocurrent data and the stratal architecture developed for the incised valley fill, this study reinterprets the trend of the axis of the incised valley to be oriented in a north-south direction. The cross section presented in this study is interpreted as a west to east transect across the eastern half of the valley fill (Fig. 15). This proposal contrasts with Van Wagoner's (1995) interpretation of a west-east trending valley in which a similar cross-section line was proposed to be parallel to the axis of a west-east oriented valley.

This new interpretation negates the need to explain why braided rivers in the Castlegate Sandstone thin and transition into low-energy fluvial systems and lacustrine settings to the east. The change in fluvial style from braided river systems in the western extent of the study area to meandering stream systems in the eastern extent can be explained by the stratal architecture of the valley fill. The eastern pinchout of the Castlegate Sandstone, therefore, can be explained as a depositional onlap along the edges of a north-south oriented valley. Facies change within the Castlegate Sandstone drainage system reflects a gradual decrease in slope and discharge as the valley filled, rather than a facies change associated with the distal edge of a terminal fan. This is a significantly different explanation for the eastern pinch out of the Castlegate Sandstone than loss of water discharge as the drainage systems flows towards the KWIS or removal of deposits by erosion.

Incised valley fills can be divided into a lower valley fill and an upper valley fill based on facies distribution. Braided-river deposits occur consistently in the lower valley fill and reflect the high energy portion of the valley fill. The lower valley fill is present exclusively in the western portion of the study area. Meandering stream deposits compose the upper valley fill. Relative sea-level rise caused a decrease in fluvial gradient, and the fluvial systems adjusted by increasing in sinuosity, forming meandering streams instead of braided rivers (Bridge, 2006).

Recent research suggests deposition of fluvial sediments occurs in stages during a relative fall in sea level: early-stage valley fill and late-stage valley fill (Wright and Marriot, 1993; Zaitlin et al., 1994; Blum and Aslan, 2006). Early-stage fill of an incised valley occurs in the deepest parts of the valley and consists of laterally amalgamated channel-fill sandstone beds with no tidal influence (Fig. 16). Deposits may show soft-sediment deformation or bank collapse due to valley-profile over-steepening (Martinsen, 1994). As relative sea level begins to rise again, the fluvial system adjusts to the change in gradient by changing its channel pattern from braided to sinuous (Schumm, 1993; Van Wagoner, 1995). Late-stage valley fill, therefore, consists of heterolithic, sinuous channel deposits that are deposited during the initial relative rise in sea level and may be tidally influenced (Martinsen, 1994; Van Wagoner, 1995; Blum and Aslan, 2006).

The portion of the Castlegate Sandstone referred to as lower valley fill in this paper is interpreted to be part of the early stage of incised valley fill (Fig. 17). The straight to slightly sinuous braided river deposits of the Castlegate Sandstone display a strong southward trend in paleocurrent direction. Because lower valley fill represents initial deposition, the main drainage direction during early Castlegate lowstand deposition was to the south.

The upper valley fill is part of the late-stage incised valley fill, and the top 10 m of the Castlegate Sandstone lowstand strata throughout the study area are upper valley-fill deposits. The eastern portion of the study area is dominated by late-stage incised valley fill. The strata are deposits of sinuous rivers that felt an increasing amount of tidal influence, corresponding to flooding of the incised valley.

The strong south-southeastern trend in paleocurrent direction in the early-stage fill of the incised valley signifies a reorientation of the drainage system that extended generally eastward from the Sevier orogenic belt into the KWIS during the previous highstand. Early-stage valley fill is present along the entire axis of the incised valley; its presence exclusively in the western portion of the study area suggests the axis of the incised valley system of the Castlegate Sandstone does not trend east-west. Instead, the axis of the valley trends north-south and parallels the axis of the foredeep and foreland basin. The southward paleoflow suggests the Castlegate Sandstone records axial drainage of the foreland basin (Fig. 18). The location of the late-stage valley fill in the updip portion of the valley is thus expected. Reinterpreting the incised valley axis to trend north-south clarifies why the eastern portion of the study area is dominated by late-stage valley fill. The late-stage valley fill onlaps the eastern margin of the incised valley.

Paleoflow recorded in the meandering stream deposits of the upper valley fill shows a significant degree of scatter. West-oriented paleoflow seen east of Sagers Canyon indicates flow was reoriented back into the foredeep, signifying a topographic high existed on the eastern edge of the foredeep. This topographic high is interpreted as the exposed foreland bulge. Paleoflow was diverted down the western flank of the exposed forebulge into the foredeep. The presence of freshwater to brackish swamp deposits and *Glossifungites* and *Scoyenia* ichnofacies overlying the Castlegate sequence boundary 2 in the eastern portion of the study area signifies a period of non-deposition and subaerial exposure of the forebulge during Castlegate lowstand deposition (Van Wagoner, 1995).

The cross-bedded oolitic ironstones at West Salt Creek, Colorado are proposed to have been deposited on the crest of the forebulge during the early stages of the Buck Tongue transgression. This interpretation places the crest of the forebulge during deposition of the Castlegate Sandstone near the present day Colorado-Utah border. The western limb of the forebulge is projected to extend as far west as Bull Canyon, where paleocurrent direction begins to show a strong western trend.

Discussion

Fluvial drainage in many active foreland basin systems is parallel to the axis of the basin. For example, the Danube River in the North Alpine Basin, the Tanana River in the Tanana Basin, the Guadalquivir River in the Guadalquivir basin, and the Ganges River in the Himalayan Basin are a few of the fluvial systems that express axial drainage of a foreland basin. Underfilled basins drain along the axis of the foredeep, or parallel to the thrust belt (Flemings and Jordan, 1989). Sediment mainly comes from the thrust belt, but erosion of an exposed forebulge will also redistribute sediment into the foredeep (Jordan, 1995; Crampton and Allen, 1995).

Because the Sevier foreland basin was inundated by the KWIS during highstand events, drainage systems generally maintained an eastward flow, away from the thrust belt into the seaway. Relative fall in sea level during lowstand events, however, could be great enough to decouple the foreland basin from the KWIS (Kamola and Hoffmeister, 2010) (Fig. 18).Based on the stratal thickness of the incised valley fill within the Castlegate Sandstone, a minimum relative sea-level drop of approximately 30 m is proposed. The relative fall was large enough to expose the forebulge and force part of the foredeep to become non-marine, isolating the foreland basin from the KWIS. The topography of the foreland basin then became the main control on lowstand drainage, reorienting paleochannels to flow south along the axis of the basin. The southward trend in paleoflow and the presence of early-stage valley fill represent flow along the axis of the foredeep, toward the proto-Gulf of Mexico.

The south-trending paleocurrents in the Castlegate Sandstone lowstand strata have been discussed in other studies. Van Wagoner (1995) proposed the changes in paleocurrent direction in the lowstand fluvial deposits resulted due to incipient Laramide deformation that formed a small anticline and locally diverted drainage. Others argued that paleocurrent trends and thickness patterns may have been influenced by pre-Cretaceous structural features such as the Pennsylvanian Uncompahgre Uplift or incipient movement of post-Castlegate Sandstone features, such as the Douglas Creek Arch (Yoshida, 1996; Miall and Arush, 2001). Yoshida et al. (1996) suggested the heterogeneity of the underlying Paradox basin may have influenced Castlegate Sandstone deposition.

Local southward drainage of Castlegate Sandstone lowstand deposits could be explained by crustal heterogeneities and incipient Laramide movement, but southward paleoflow is common to lowstand deposits throughout the KWIS (Slingerland and Keen, 1999; Table 1). The widespread drainage patterns in the lowstand deposits indicate a regional feature was responsible for flow diversion. The forebulge is a likely candidate for that regional feature, suggesting the dominant control on Castlegate Sandstone lowstand deposition was the topography of the foreland basin.

Westward flow direction of the Castlegate Sandstone lowstand strata has not been recognized prior to this study. Paleoflow indicators in between Strychnine Wash and Bull Canyon show a general trend to the west. Westward drainage confirms a significant topographic high was present in the Sevier foreland basin, east of the foredeep (Fig. 19). Channel-fill deposits containing westward flow indicators are thin, tidally influenced late-stage valley-fill strata, and they are interpreted to record erosion of the forebulge. Erosion of the forebulge would occur only if the forebulge surface lay near or above sea level (Jordan and Flemings, 1991; Crampton and Allen, 1995), as significant forebulge erosion is difficult under marine conditions (Crampton and Allen, 1995). The presence of non-marine facies, like the freshwater to brackish-water palynomorph assemblages found in the eastern portion of the study area, are interpreted as further evidence that the forebulge was partially to wholly exposed. Deposition of the swamp facies occurred during the last stage of lowstand deposition of the Castlegate Sandstone strata.

The ooids in the eastern portion of the study area are interpreted to have been deposited at the crest of the forebulge during the early stages of the transgression associated with the Buck Tongue. Localized ooidal ironstones, up to 2 m thick, have previously been recognized to be associated with forebulges and topographic highs (Plint et al. 1993; Taylor and Curtis 1995; Donaldson et al. 1999; Taylor et al., 2002). Although the ooidal ironstones have not been traced north into the subsurface and Castlegate strata are eroded to the south, Van Wagoner (1995) interprets the ooidal ironstones near West Salt Creek as a part of a thin, north-northwest trending belt. While the belt of ooidal ironstones is interpreted by Van Wagoner (1995) to represent the Castlegate lowstand shoreline, it has since been interpreted to be the crest of the Castlegate forebulge (Yoshida et al., 1998; Miall and Arush, 2001; Taylor et al., 2002).

Change in paleocurrent direction from south to west reflects the change from foredeep- and forebulge-dominated flow to forebulge-dominated flow that occurs near Sagers Canyon. Previous studies have yet to definitively calculate the width and height of the forebulge, partially because forebulges are positive migratory features and are preserved mainly as erosional surfaces (Jacobi, 1981). By measuring from where the foredeep is interpreted to transition into the forebulge to the location of the ooidal ironstone, the western arm of the forebulge is proposed to be 47 km (29 mi) wide. The height of the forebulge is not determined.

Facies changes and paleoflow data collected in this study place the forebulge on the Utah-Colorado border around 80 to 78 my. DeCelles and Currie (1996) interpret the overall movement of the foreland basin from central Utah to eastern Utah to occur at a rate of ~0.5 cm/yr, which would place the forebulge crest in western Colorado during the Campanian, with a flank extending into eastern Utah (Miall and Arush, 2001). Any crustal heterogeneity may have affected the rate of forebulge migration; forebulge migration may have been transiently fixed on a pre-existing zone of crustal weakness (Waschbusch and Royden, 1992) and may not have migrated with the shifting thrust and sediment loads (Jordan, 1981).

To accept the influence of the forebulge during deposition of the Castlegate Sandstone, it is important to recognize that the Sevier foreland basin was underfilled during part of the Campanian. Based on isopachs of the non-marine Cretaceous strata, the Sevier foreland basin is interpreted to be overfilled by the Cenomanian-Turonian (Jordan, 1981). Overfilled basins exhibit traverse drainage instead of axial drainage, and sediment can be transferred outside the range of tectonic influence (Jordan, 1995). If the Late Cretaceous Sevier foreland basin were overfilled, the forebulge would be buried, preventing Castlegate lowstand drainage systems from expressing axial drainage (Fig. 20). The fact that the Castlegate appears to exhibit axial drainage suggests the basin was actually in an underfilled state. Yang (2010) modeled the possibility of tectonically driven underfilled-overfilled cycles in the northern Cordilleran basin, focusing on the presence of the forebulge and its influence within the basin. Episodic thrusting in the orogenic wedge and low thrust velocity (about 2mm/yr in Utah) means there was a relatively long hiatus between major thrusting events (Yang, 2010). During hiatuses in thrusting, denudation filled the foredeep with sediment shed from the orogen and created overfilled conditions in the basin. When a new thrusting event drives the basin to return to an underfilled state, it completes an underfilled-overfilled cycle (Yang, 2010). Although the Sevier foreland basin was overfilled during the Cenomanian (Jordan, 1981), it was underfilled during the Campanian. Preserved fresh-water swamp deposits establish that the forebulge was subaerially exposed, and the southern trend of paleoflow show the Castlegate Sandstone lowstand fluvial systems expressed axial drainage.

Conclusions

Analysis of paleocurrent directions and facies changes of the incised valley fill of the Castlegate Sandstone led to a reinterpretation that the incised valley trended north-south. With the understanding that the lowstand deposits are part of a north-south valley system, the facies changes are reinterpreted to represent stages of valley fill rather than a down-dip thinning resulting from the loss of water discharge as the drainage systems flow toward the KWIS.

Early-stage valley fill was differentiated from late-stage valley fill, and the location of these stages of valley fill was evaluated. Early-stage valley fill of the Castlegate Sandstone comprises thick, multistory channels deposits in which paleocurrents express a south-southeastward trend with minor data dispersion, which signifies minor variation in flow direction. Early-stage fill is found only in the western portion of the study area, as flow during this time was concentrated in the foredeep of the foreland basin.

Early-stage valley fill deposits were overlain by late-stage valley fill. Deposition within the upper portion of the Castlegate incised valley fill is limited to single-story channel-fill deposits. Deposition occurred in sinuous, meandering systems. Paleocurrents display a wider degree of scatter than the early-stage fill but show an overall trend to the south.

The Castlegate Sandstone in eastern portion of the study area is composed entirely of upper valley fill. The meandering stream deposits show abundant lateral accretion and tidal influence. Deposition of these fluvial systems occurred during the latest stages of valley fill. The deposits display paleoflow to the west, reflecting channel flow from the exposed forebulge into the foredeep.

The forebulge was exposed during a relative fall in sea level that also forced the foredeep to become non-marine within the study area. Subaerial exposure of the forebulge decoupled the foredeep from the KWIS, allowing the topography of the foreland basin to dominate drainage systems. Free from the influence of the KWIS, the Sevier foreland basin exhibited axial drainage in a manner typical of most modern subaerially exposed foreland basins.

The existence of freshwater swamp facies and overlying ooidal ironstones helps isolate the location of the forebulge during deposition of the Castlegate Sandstone. The thin belt of ooidal ironstones is thought to be deposits that formed on the crest of the forebulge during the transgression of the Buck Tongue. The location of channels displaying westward drainage helps define the width of the western limb of the forebulge. The crest of the forebulge is located nearly 20 km east of the Colorado-Utah border, and fluvial systems drain the western limb of the forebulge from Horse Pastures to Sagers Canyon into the foredeep.

The results of this study suggest that paleoflow recorded in lowstand strata of the Castlegate Sandstone represents axial drainage within the foreland basin. Axial drainage of the Sevier foreland basin was restricted to periods of lowstand conditions, during which the basin topography influenced the drainage system. This interpretation of the Castlegate Sandstone lowstand strata stresses the importance of recognizing the impact of the foreland basin on drainage patterns during their deposition. A better understanding of the axis of incised valleys may lead to new interpretations relating to the stratal evolution of the foreland basin and a better understanding of the complex interplay between the KWIS and the foreland basin especially during lowstand events.

Figures

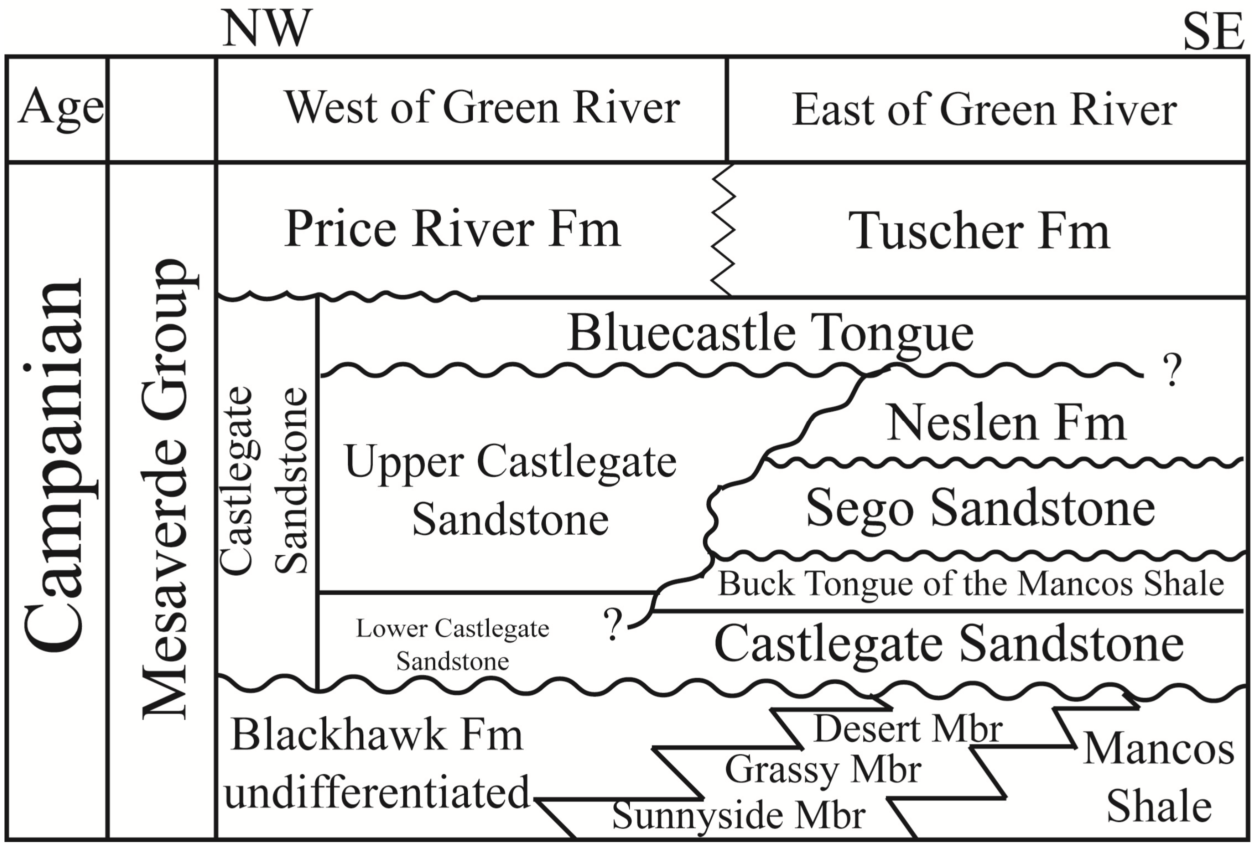

Figure 1: Stratigraphic column of the Mesaverde Group illustrating the difference in nomenclature of lithostratigraphic units in the Book Cliffs based on location. Fm = formation; Mbr = member. Adapted from Fouch et al., 1983 and Van Wagoner, 1995.

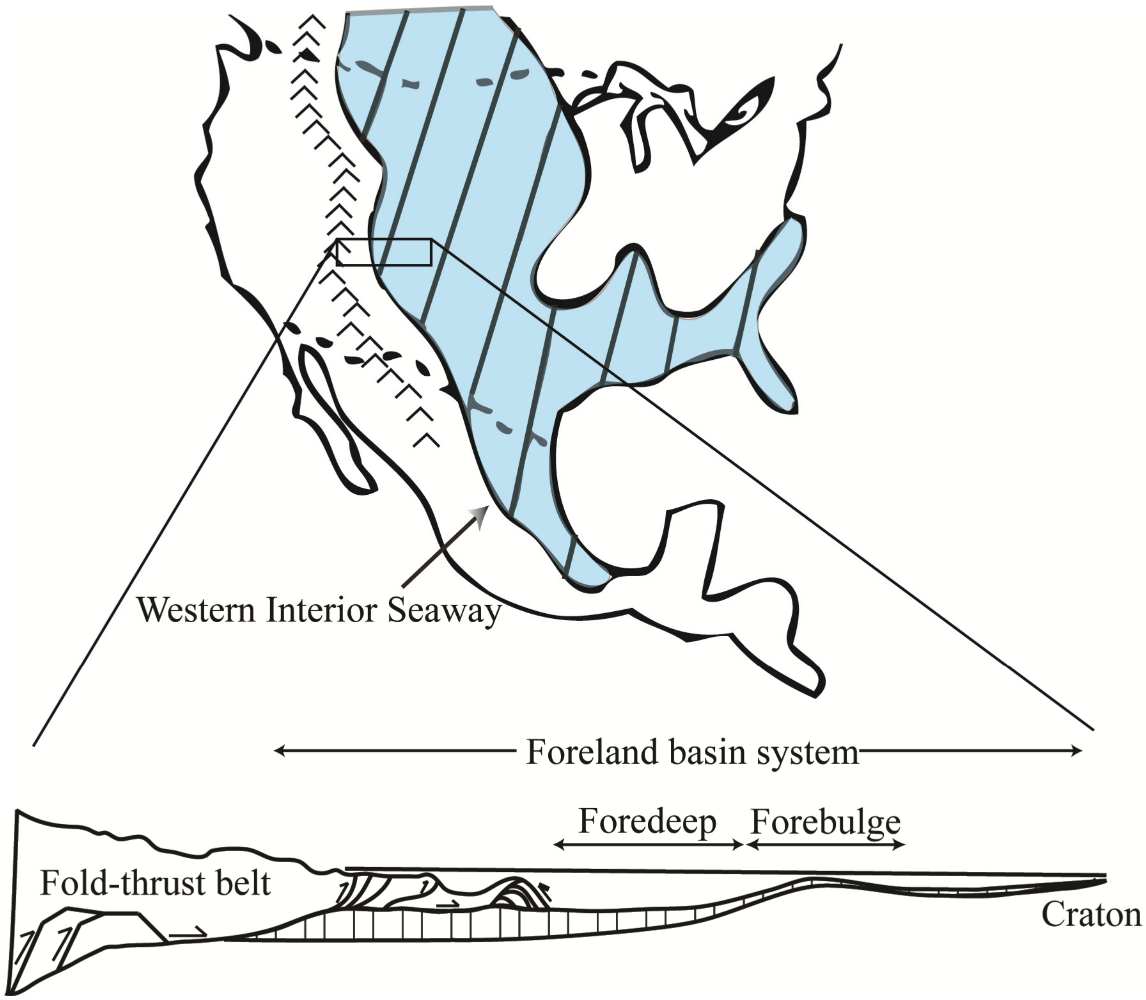

Figure 2: During the Late Cretaceous, a large portion of North America was covered by an epeiric sea known as the KWIS. The foreland basin sat on the western edge of the KWIS. The cross section shows a typical foreland basin system.

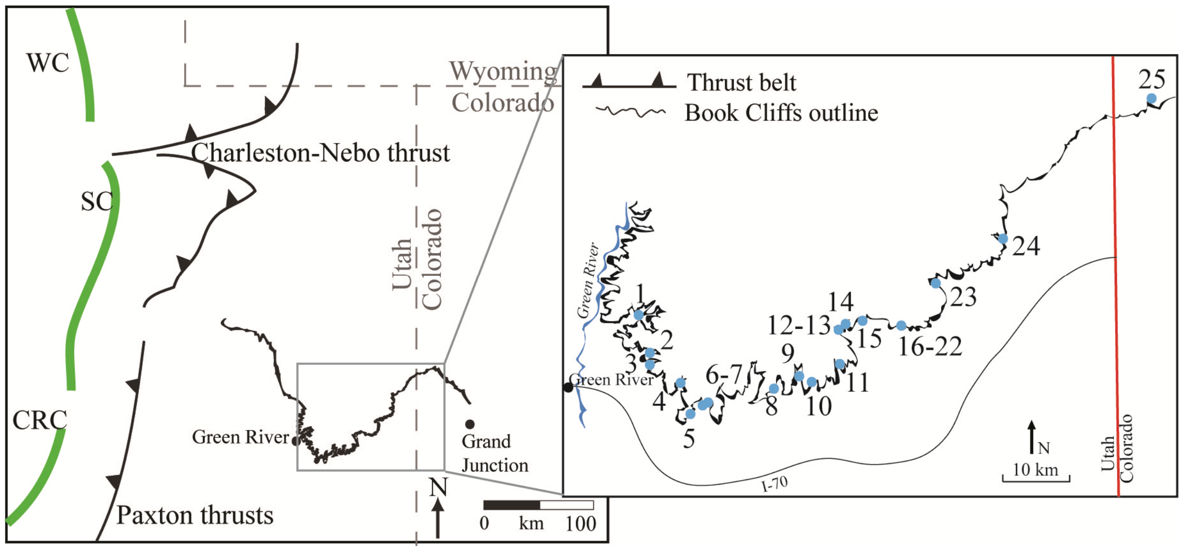

Figure 3: The study area is in the southern portion of the Book Cliffs, Utah, U.S.A. CRT= Canyon Range thrust, PT= Pavant thrust. The thick gray lines represent the Wasatch Culmination (WC), Santiquin Culmination (SC) (potential sediment source), and the Canyon Range Culmination (CRC). Edited from Horton et al. (2004), DeCelles and Coogan (2006), and Hampson (2010).

Gray dots represent locations of canyons measured sections: 1) Tuscher Canyon; 2) Coal Canyon; 3) Stubb Canyon; 4) Horse Canyon; 5) Hatch Mesa; 6-7) Floy Canyon; 8) Blaze Canyon; 9) Thompson Canyon; 10) Jeep Trail; 11) Sagers Canyon; 12-13) Bull Canyon; 14) Calf Canyon; 15)Strychnine Wash; 16-22) Horse Pastures; 23) Big Hole Wash; 14) Sulpher Canyon; 25) West Salt Creek.

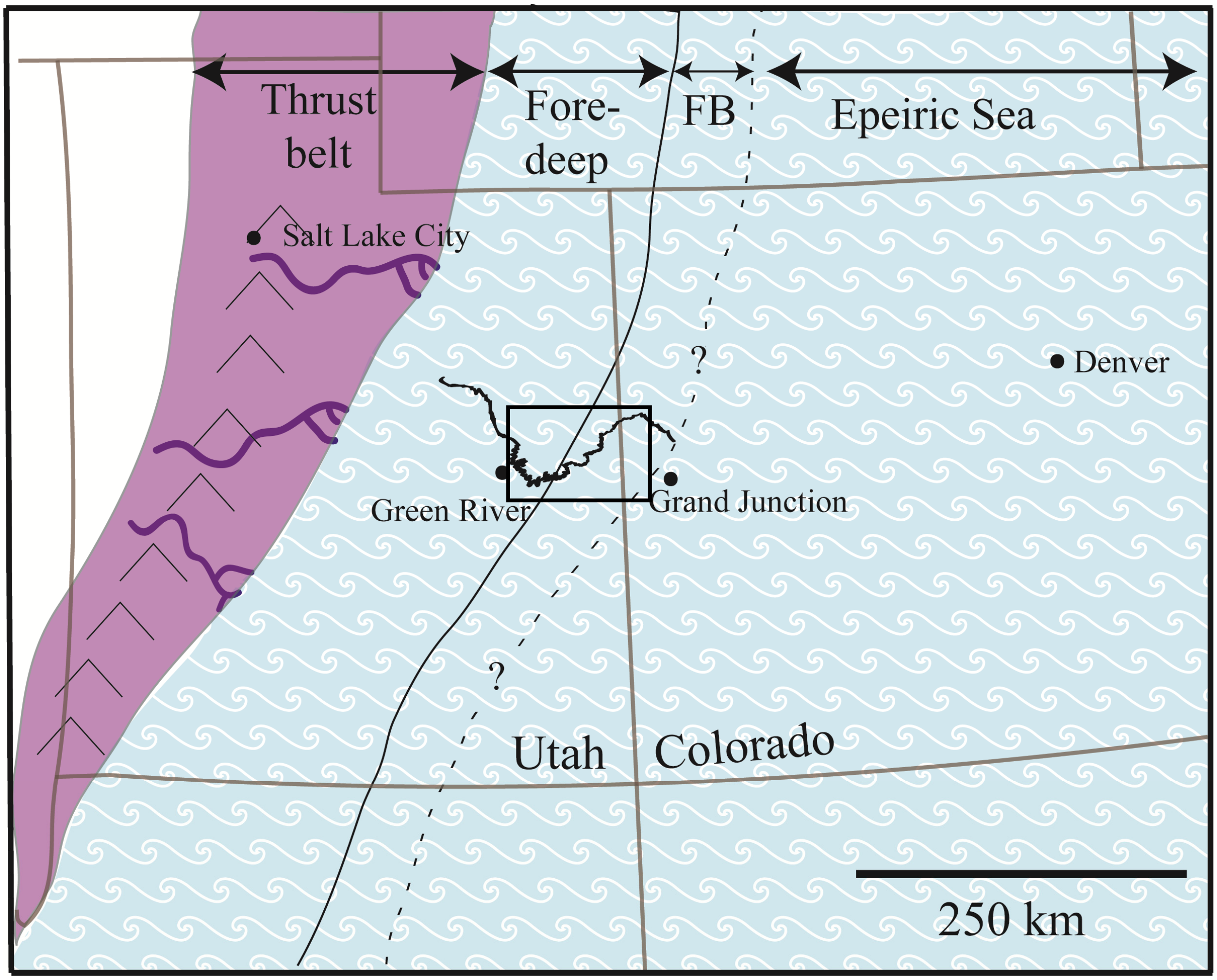

Figure 4: Schematic of the Sevier foreland basin in highstand conditions during the Campanian. The black line outlines the Book Cliffs. FB= forebulge. The black box outlines the study area.

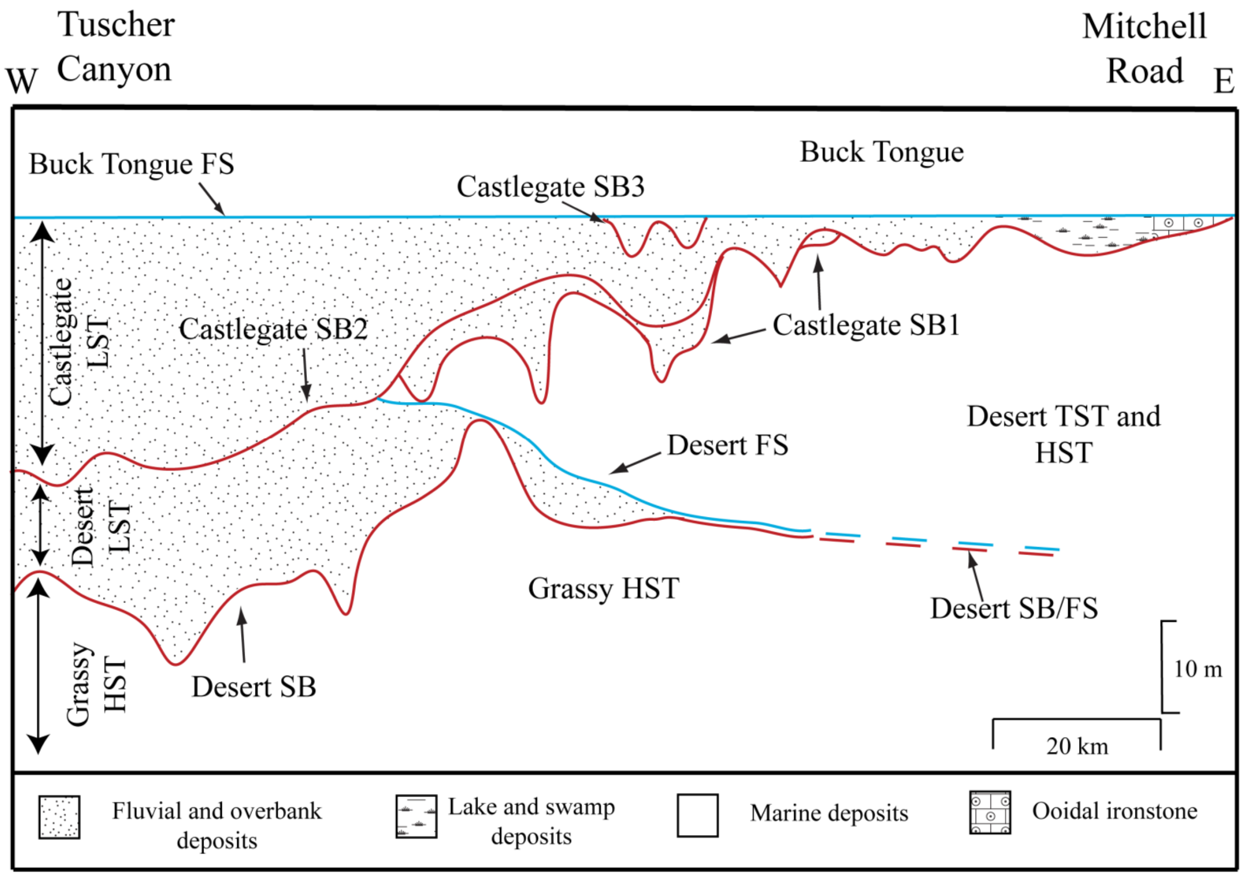

Figure 5: Sequence stratigraphy of the Desert Member of the Blackhawk Formation and the Castlegate Sandstone in the southern Book Cliffs as interpreted by Van Wagoner (1995). SB = sequence boundary; FS = flooding surface; LST = lowstand systems tract; TST = Transgressive systems tract; HST = highstand systems tract.

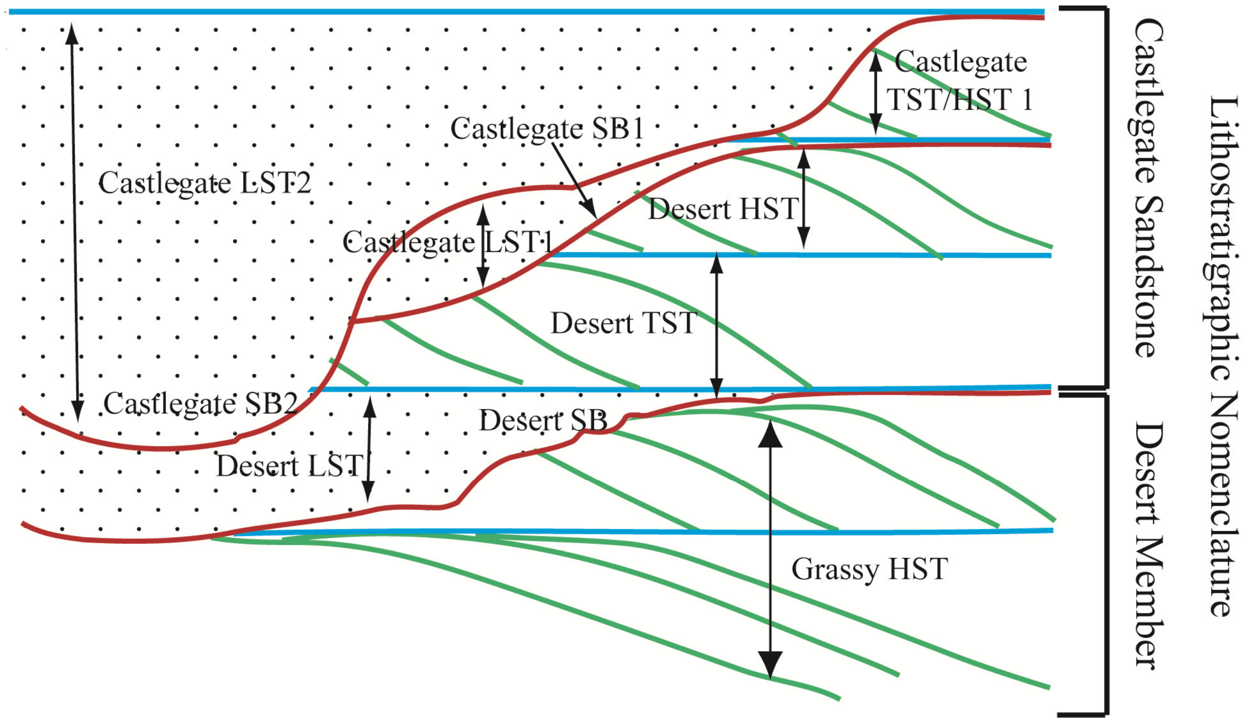

Figure 6: Sequence stratigraphic interpretation of the Desert Member and Castlegate Sandstone, modified from Van Wagoner (1995). The thick gray lines represent parasequence boundaries; thin gray lines represent progradational packages of strata. The lithostratigraphic nomenclature does not correlate to the sequence stratigraphic units.

Figure 7: Tabular cross bedding within the braided stream facies. The black lines on the Jacob staff are separated by 10 cm. The bottom bed is 20 cm thick.

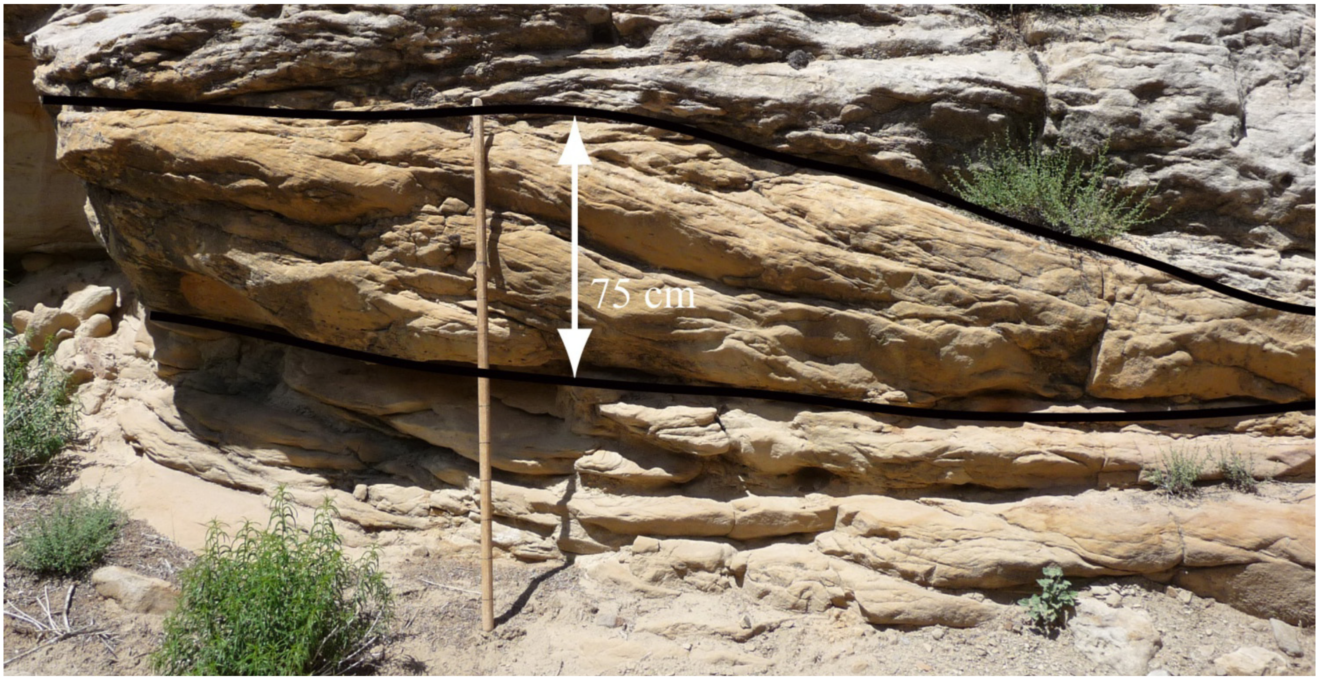

Figure 8: Large, sigmoidal- to wedge-shaped bars are part of the braided stream facies (Facies two). The Jacob staff measures 1.5 m long and the bar is 75 cm thick.

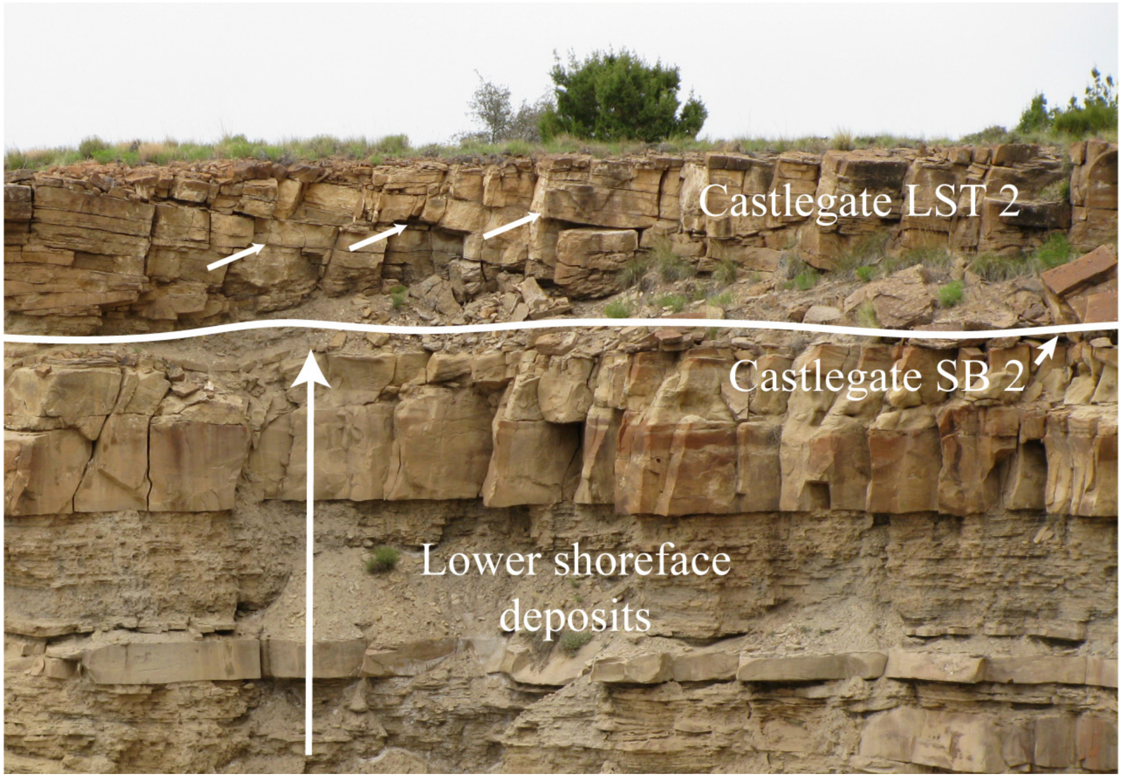

Figure 9: The short white arrows mark laterally accreting beds of channel deposits in Horse Pastures, UT. The large white arrow shows upward coarsening package of lower shoreface deposits. The Castlegate Sandstone lowstand systems tract (LST) 2 overlies the Castlegate sequence boundary (SB) 2 and consists of a fining upward single story channel that incises into the lower shoreface hummocky cross stratified beds. The top of the Castlegate LST 2 correlates to the Buck Tongue flooding surface.

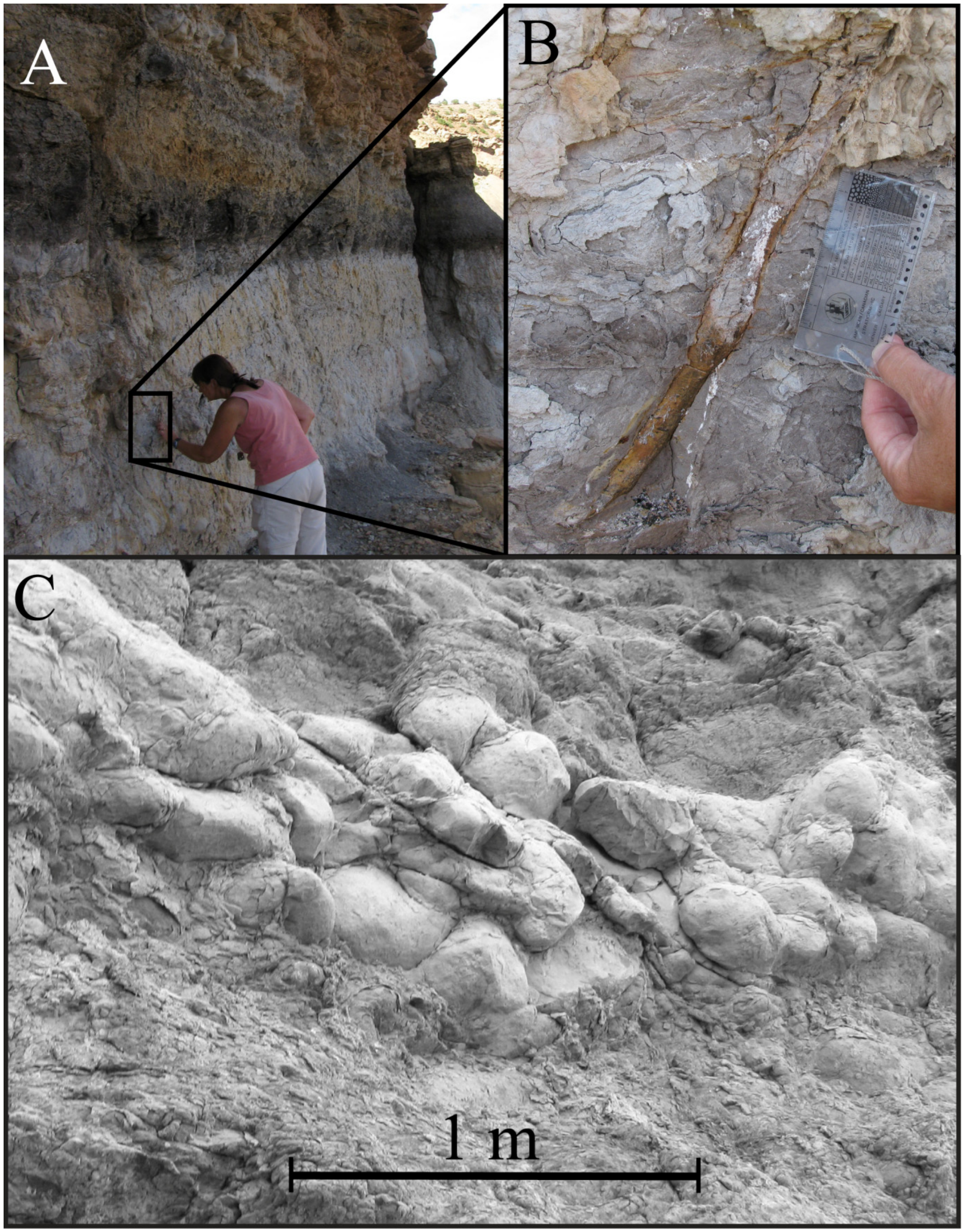

Figure 10: Thick, light gray to dark gray rooted siltstone (A, B) and mottled surfaces (C) are part of a paleosol near the top of the incised valley fill in Floy Canyon. The thick siltstone can be traced laterally until they are replaced by single story meandering channels.

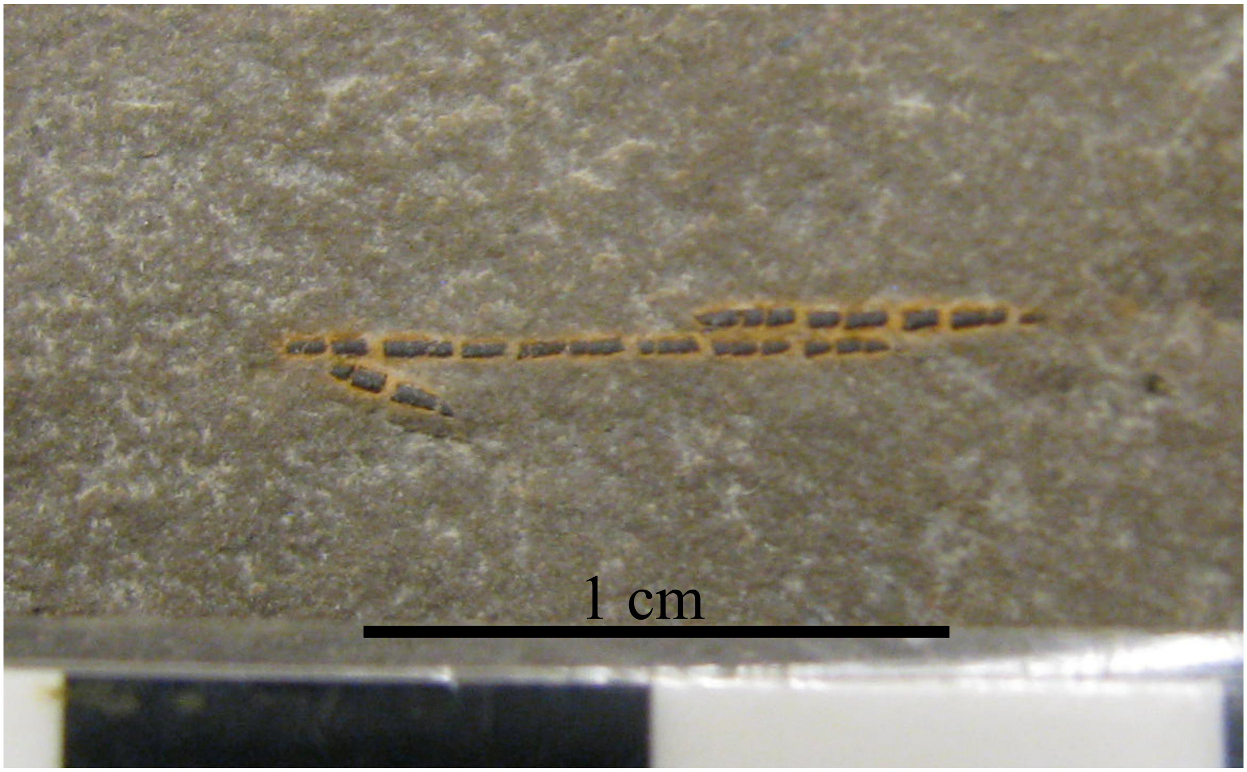

Figure 11: *Equisetales* found in the thin gray siltstone above the Castlegate sequence boundary 2 in the eastern portion of the study area. *Equisetales* are common in wetlands and marshlands.

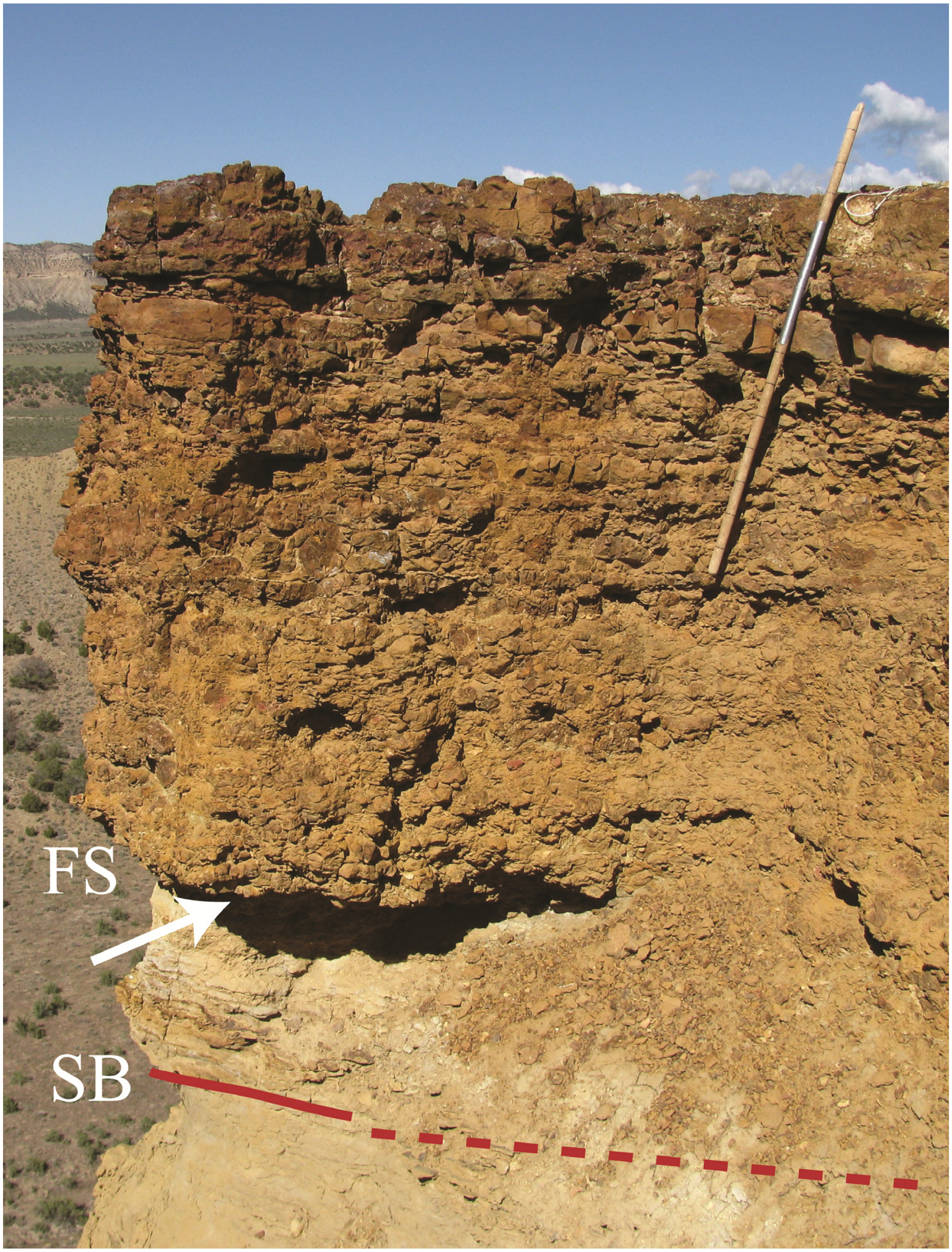

Figure 12: The Castlegate sequence boundary 2 (SB) located at West Salt Creek, Colorado is overlain by 50 cm of silty sandstone which contain abundant organic matter interpreted as swamp deposits. The swamp deposits are overlain by two meters of ooidal ironstone. The base of the ooidal ironstone represents the initial flooding surface (FS) of the Buck Tongue of the Mancos Shale. The Jacob staff in the upper right corner of the picture is 1.5 m long.

Canyon	Paleocurrent Trend	Paleocurrents Gathered
Tuscher Canyon	168^o	48
Coal Canyon	151^o	35
Stubb Canyon	152^o	39
Horse Canyon	171^o	39
Hatch Mesa	182^o	6
Floy Canyon	194^o	35
Blaze Canyon	160^o	33
Thompson Canyon	153^o	42
Jeep Trail	174^o	29
Sagers Canyon	201^o	12
Bull Canyon	293^o	9
Strychnine Wash	170^o	12
Horse Pastures	220^o	35

Table 1: Average trend of paleocurrents gathered in each canyon. The names of the canyons can be associated with their locations in Figure 3.

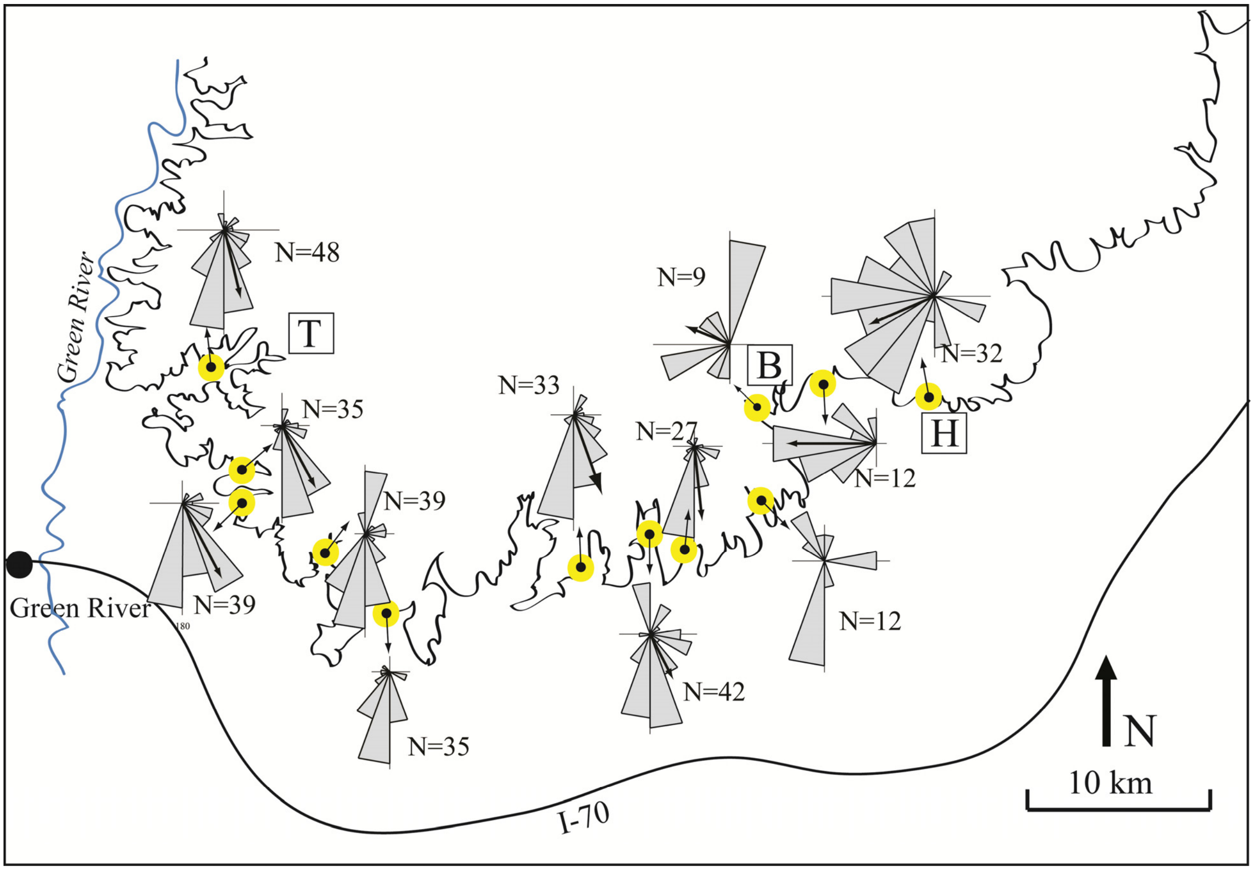

Figure 13: Paleocurrent directions collected between Tuscher (T) and Bull Canyons (B) show predominant flow direction south (arrows represent direction and magnitude of the trend). Paleocurrent directions collected between Bull Canyon and Horse Pastures (H) indicate a westward flow.

Figure 14: Stacked fluvial channels in the western portion of the study area. Each white arrow marks a bedset. The cliff face is 40 m high.

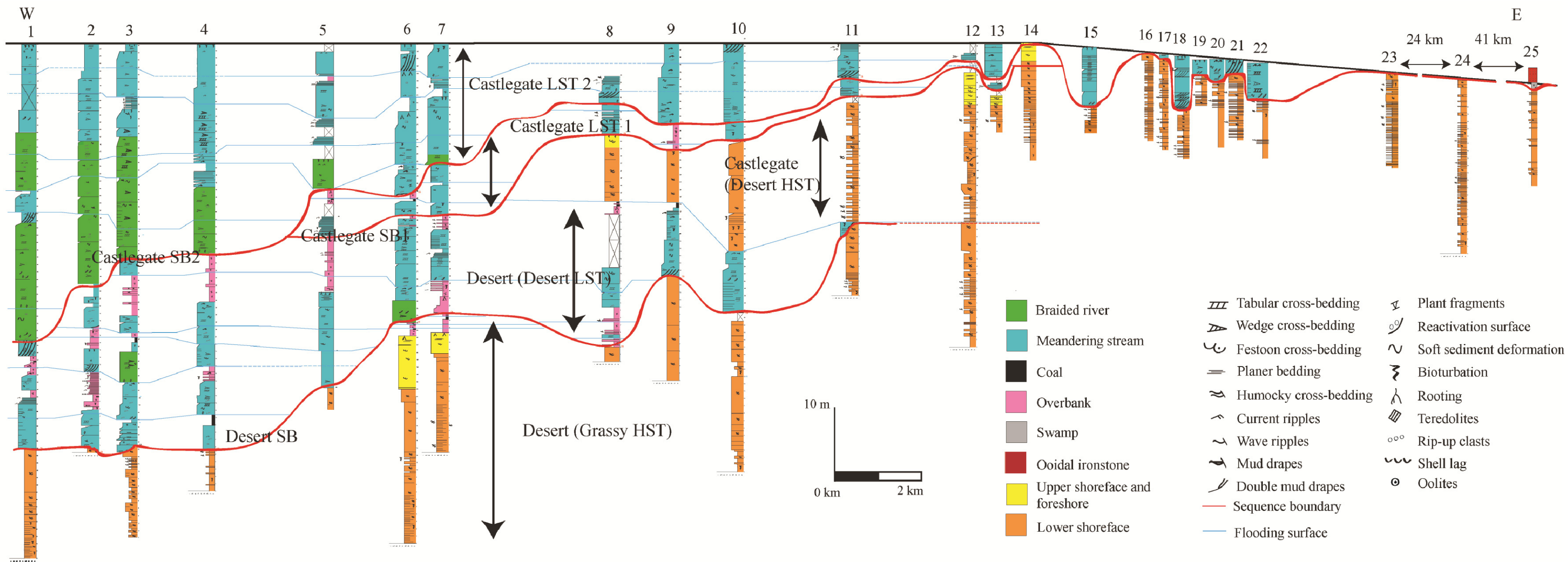

Figure 15: The cross section represents a west-east transect across part of the width of the incised valley. The axis of the Castlegate incised valley fill is in and out of the page, and the valley margin is seen climbing to the east. 1: Tuscher Canyon; 2: Coal Canyon; 3: Stubb Canyon; 4: Horse Canyon; 5: Hatch Mesa; 6-7: Floy Canyon; 8: Blaze Canyon; 9: Thompson Canyon; 10: Jeep Trail; 11: Sagers Canyon; 12-13: Bull Canyon; 14: Calf Canyon; 15: Strychnine Wash; 16-22: Horse Pastures; 23: Big Hole Wash; 24: Sulpher Canyon; 25: West Salt Creek.

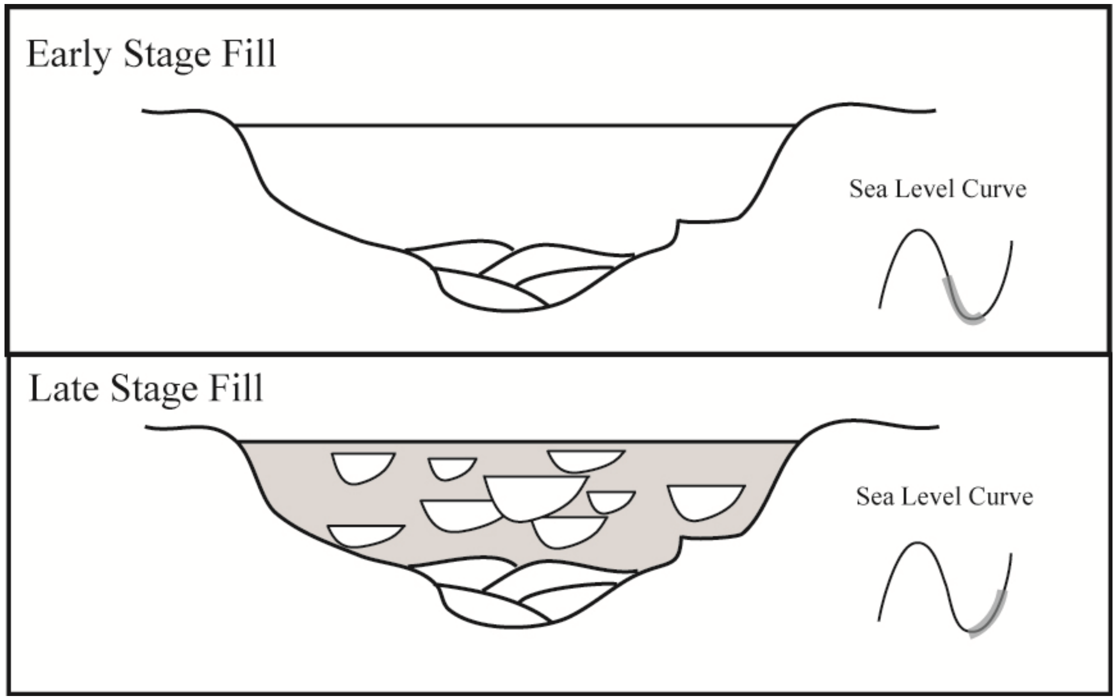

Figure 16: Conceptual diagram comparing early-stage valley fill to late-stage valley fill. Early-stage fill consists of stacked braided stream deposits with little overbank preservation while late-stage fill is composed of tidally influenced heterolithic, sinuous channel fill deposited during the initial relative rise in sea level. Modified from Martinsen (1994).

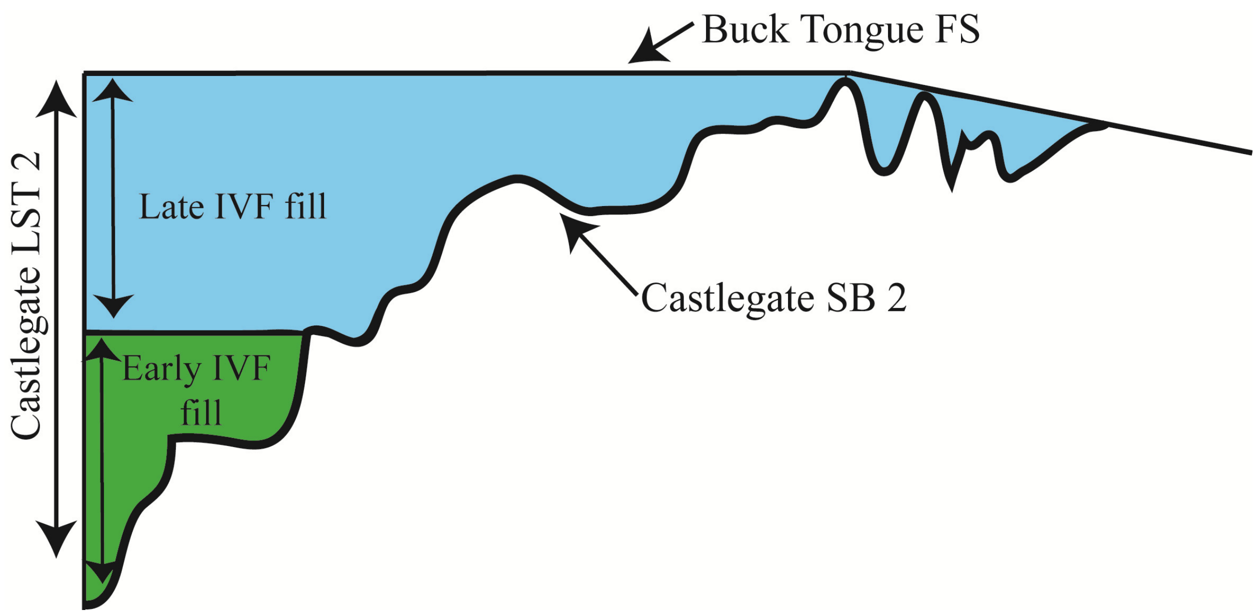

Figure 17: Early-stage incised valley fill (IVF) is concentrated in the western portion of the study area, close to the sequence boundary (SB). Late-stage valley fill was deposited above the early-stage valley fill deposits in the western portion of the study area and makes up the sole deposits to the east.

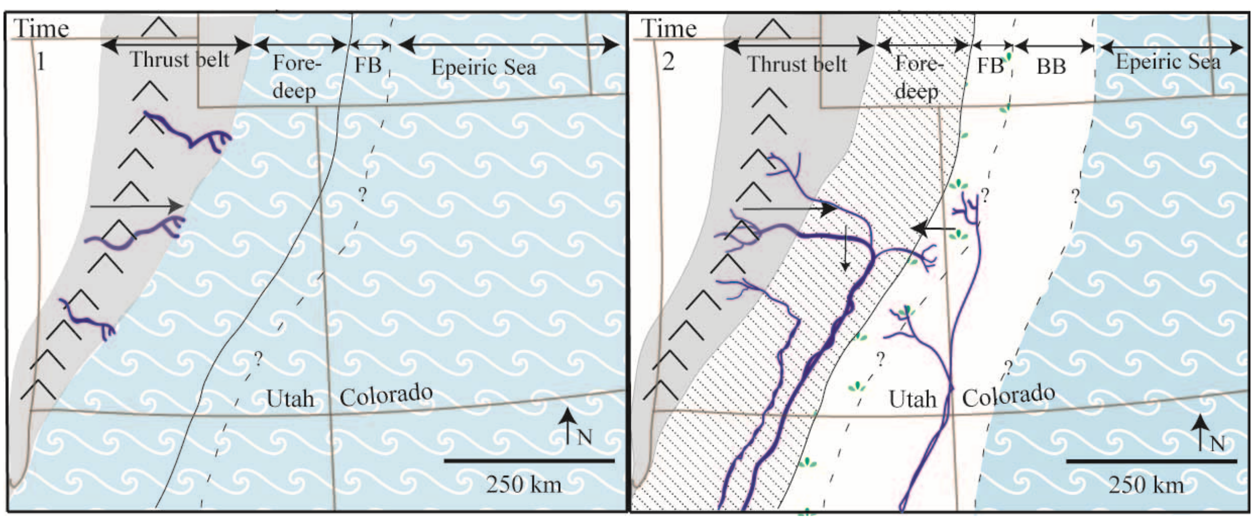

Figure 18: When relative sea level was high (Time 1), the foredeep and forebulge (FB) was inundated. Drainage systems from the Sevier thrust belt flowed east into the KWIS. During lowstand events (Time 2) the forebulge was exposed and drainage was to the south along the axis of the foredeep. The location of the western shoreline of the epeiric sea is not identified. BB= backbulge.

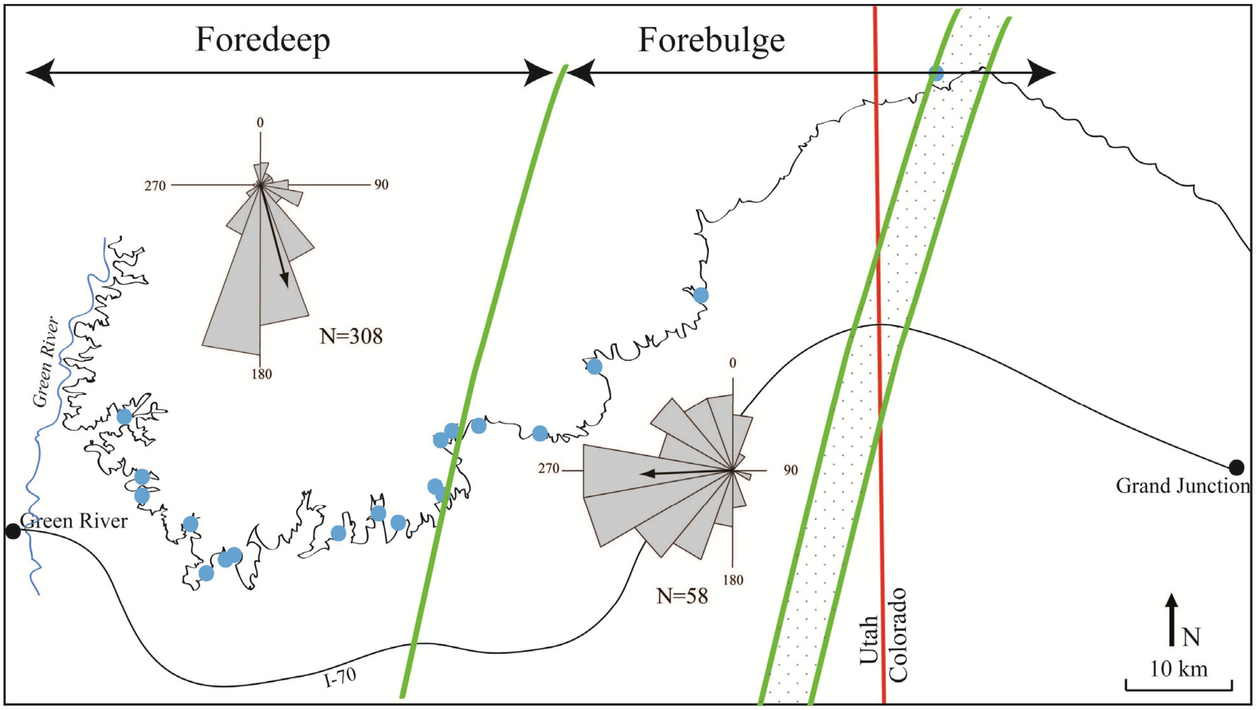

Figure 19: Outline of the southern Book Cliffs showing the western extent of forebulge to crest of the forebulge (stippled pattern). The crest of the forebulge is projected south of West Salt Creek based on the projected oolite belt of Van Wagoner(1995). Paleocurrent data taken from deposits found in the foredeep show a south-southeast trend. Paleocurrent data in the eastern part of the study area display a western flow, and are used to identify the location of the western limb of the forebulge.

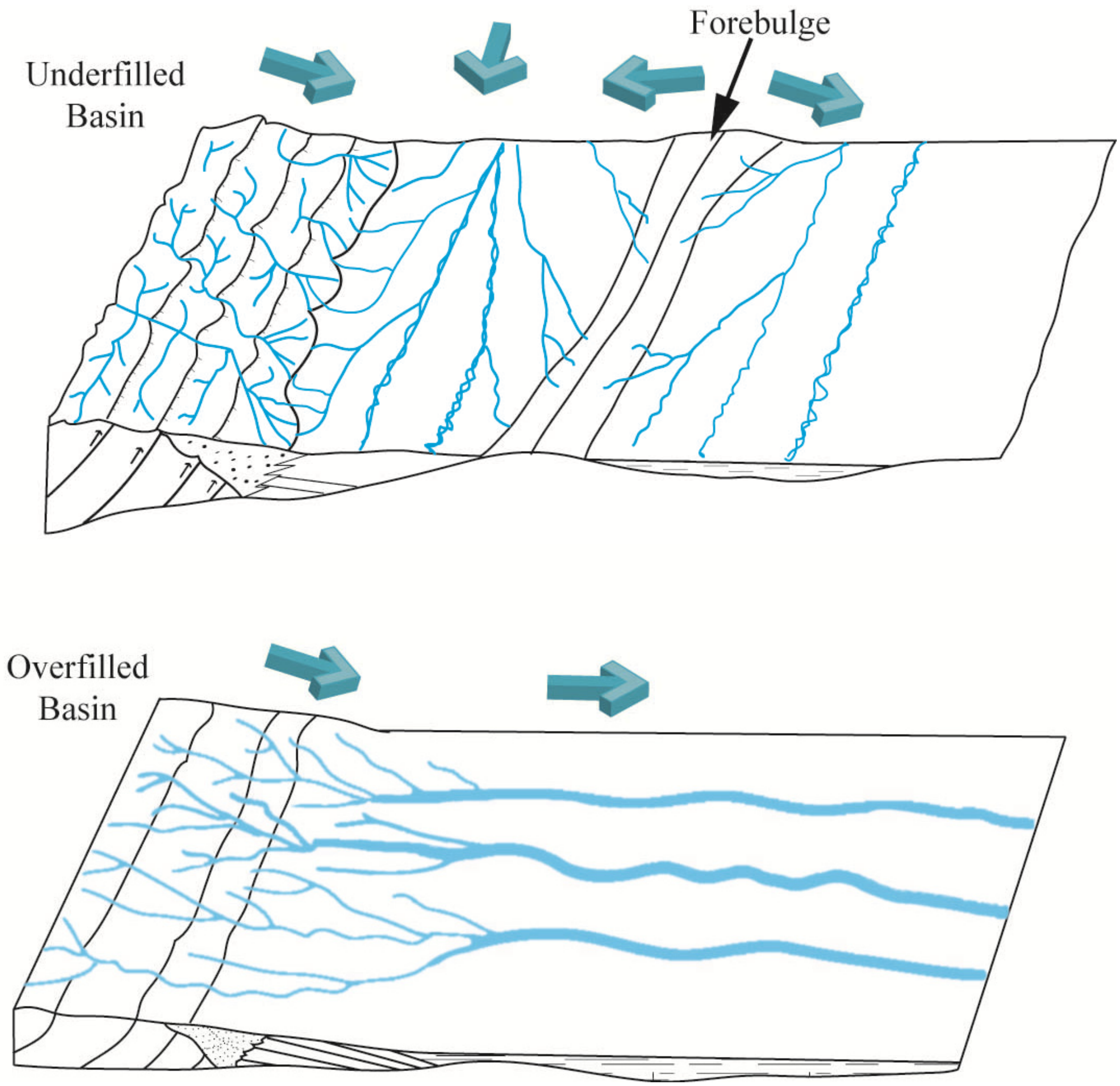

Figure 20: In an underfilled basin, drainage comes down from the thrust belt into the foredeep and drains axially. The forebulge is a topographic high that sources drainage into the foredeep and also towards the craton. In overfilled basin conditions the sediment shed from the thrust belt have shifted the subsidence of the basin, burying the forebulge and leading to transverse drainage.

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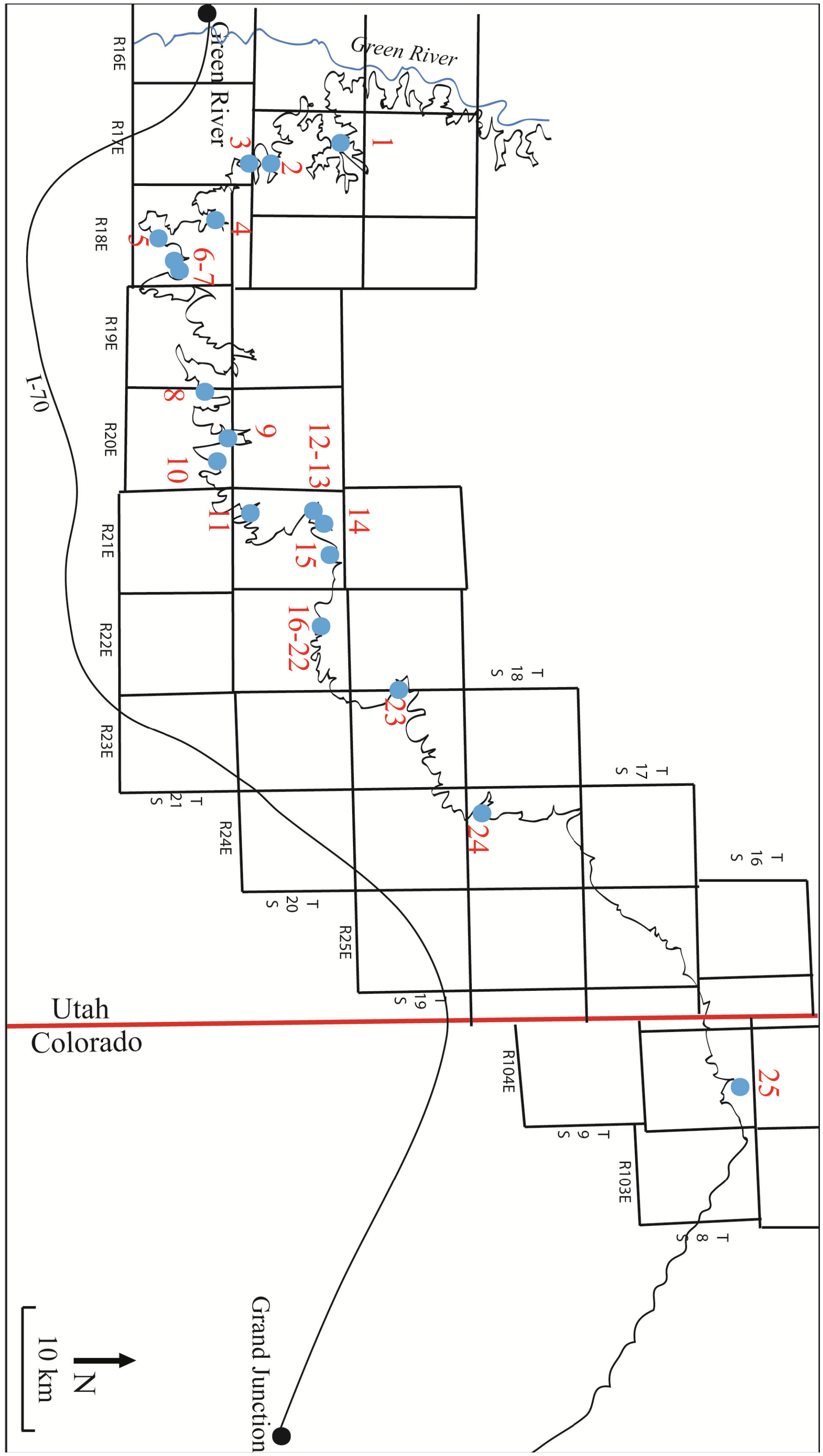
APPENDIX A: Townships and Ranges of Measured Sections

1: Tuscher Canyon, Sec. 8 T20S R17E
2: Coal Canyon, Sec. 28 T20S R17E
3: Stubb Canyon, Sec. 34 T20S R17E
4: Horse Canyon, Sec. 9 T21S R18E
5: Hatch Mesa, Sec. 27 T21S R18E
6-7: Floy Canyon, Sec. 23 T21S R18E
8: Blaze Canyon, Sec. 7 T21S R20E
9: Thompson Canyon, Sec. 10 T21S R20E
10: Jeep Trail, Sec. 2 T21S R20E
11: Sagers Canyon, Sec. 5 T21S R21E
12-13: Bull Canyon, Sec. 4 T20S R21E
14: Calf Canyon, Sec. 3 T20s R21E
15: Strychnine Wash, Sec. 2 T20S R21E
16-22: Horse Pastures, Sec. 6 T20S R22E
23: Big Hole Wash, Sec. 9 T20S R22E
24: Sulpher Canyon, Sec. 32 T18S R24E
25: West Salt Creek, Sec. 27 T7S R104W


Green River
Green River
I-70
Utah
Colorado
Grand Junction
N
10 km
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R17E
R18E
R19E
R20E
R21E
R22E
R23E
R24E
R25E
R104E
R103E
T 16 S
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T 18 S
T 19 S
T 20 S
T 21 S
T 8 S
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APPENDIX B: Ooidal Ironstone

The presence of ooidal ironstone in the eastern part of the study area is somewhat unexpected in the siliciclastic dominated basin. Marine grains primarily form on shallow, clastic-starved shelves (Macquaker et al., 1996). Ooidal ironstones are present in the eastern most outcrop of the Castlegate Sandstone, near West Salt Creek, Colorado and Mitchell Road, Colorado (Van Wagoner, 1995; Taylor et al., 2002) (Fig. 1). Jurassic ooidal ironstones are locally present in the Book Cliffs, near Hatch Mesa (Chan, 1992). The ooidal ironstones that make up the most distal Castlegate outcrops have been extrapolated to form linear bars that run generally north-south and are interpreted as a down-dip expression of the Castlegate Sandstone lowstand shoreline (Van Wagoner, 1995). These localized, linear bodies of ooidal ironstone are reinterpreted as the crest of the forebulge during Castlegate Sandstone lowstand deposition.

The West Salt Creek outcrop consists of several meters of thin beds of hummocky cross stratified sandstone that are overlain by gray, silty sandstone with palynomorphs of vascular plant fragments, such as preserved *Equisetales*. Modern *Equisetales*, commonly called horsetail or joint grass, has jointed stems, whorled leaves, and branches with a terminal cone on fertile shoots and is found in swamps. The erosional surface correlates updip to the Castlegate sequence boundary 2 and the siltstone is swamp deposits (see Chapter 2). Overlying the swamp deposits are two meters of vaguely crossbedded ooidal ironstones.

The ooidal ironstone is composed of fine to medium quartz grains and well sorted ooids in a matrix dark brown to red siderite and clay. While most of the ooids have been replaced by ferroan dolomite, some ooids retain their original fabric (Taylor et al., 2002). The ooids are characterized by berthierine or kaolinite interlaminated with goethite and siderite cement

(Donaldson et al., 1999; Taylor et al., 2002). Nuclei of the ooids are quartz grains, feldspar grains, and shell fragments.

The ooids formed in high energy, shallow water environments along the shoreline during a period of decreased siliciclastic input (Van Wagoner, 1995; Taylor et al., 2002). Periodic exhumation by storms concentrated the grains in a high energy oxic environment, and as the storm energy waned and transitioned from the oxic zone to a suboxic zone the ooids developed new coatings (Taylor and Curtis, 1995; Macquaker et al., 1996; Taylor et al., 2002). The dolomitization of the ooids was a diagenetic event that is possibly related to the flooding during Buck Tongue deposition as laterally persistent dolomite cements are present throughout the flooding surface beneath Buck Tongue deposits (Taylor et al., 2000).

The ooidal ironstone at West Salt Creek has fewer allochems and more mud in its matrix than the ooidal ironstone present at Mitchell Road. The abundance of allochems and lack of mud content at Mitchell Road represents a higher energy environment, suggesting storm energy was stronger to the east (Taylor et al., 2002).

Localized ooidal ironstones have been associated with forebulges and topographic highs such as the Bad Heart Formation of the Western Canada foreland basin (Donaldson et al., 1999), the Paddy Member Ironstone of the Peace River Formation (Taylor 2002), and the Lower Kaskapau Formation (Plint, 2000). Deposition of these shallow-water and iron-rich facies appears to have been localized over the crest and flanks of a forebulge (Donaldson et al., 1999) and express morphology consistent with winnowing of clastic input over a swell (Plint, 2000). The Castlegate ooidal iron stones have been interpreted as a possible crest of the Castlegate forebulge (Yoshida et al., 1998; Miall and Arush, 2001; Taylor et al., 2002).

The timing of ooidal ironstone deposition has been suggested to be during the latest stages of regression, along sequence boundaries, due to sediment trapped up dip (Van Wagoner, 1995), or due to sediment bypass (Macquaker et al., 1996). Topographic highs can exert a strong hydrodynamic force that moves clastic material around the structure, instead of onto it (Taylor and Curtis, 1995), providing enough energy to form ooidal shoals. More likely, however, is that ooid formation occurred during the earliest stages of transgression, when sea level was still low enough to be partially oxic but the clastic shelves were sediment starved (Chan, 1992; Donaldson et al., 1999; Taylor et al., 2002).

The presence of ooidal ironstones indicates the existence of the forebulge in western Colorado. Ooid formation occurred during the early stages of Buck Tongue transgression when sea level was slowly rising, trapping sediment closer to the Sevier thrust belt. The ooids developed on the crest of the forebulge, the highest topographic point, where they were subject to reworking and growth due to the waxing and waning energy of storms. Higher allochem ratios and lower mud ratio in the Mitchell Road outcrop suggest storms came from the east indicating the forebulge, when inundated, no longer had an effect on the circulation patterns of the Cretaceous Western Interior Seaway.

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Figures

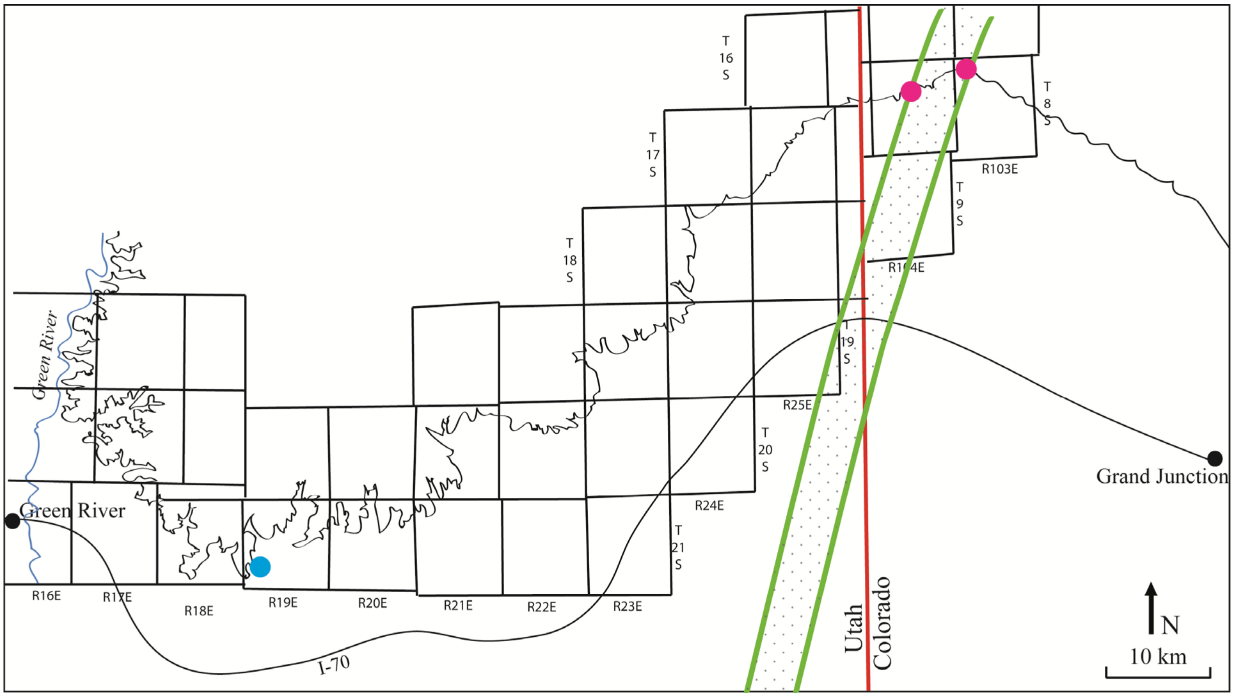


Figure 1: Location of ooidal ironstone in the Book Cliffs, Utah. Pink dots represent outcrops of Cretaceous ooidal ironstone and the blue dot is the location of the Jurassic ooidal ironstone. The green lines represent the extrapolated belt of ooids (Van Wagoner, 1995).

APPENDIX C: Measured Section

The measured sections are provided in alphabetical order.

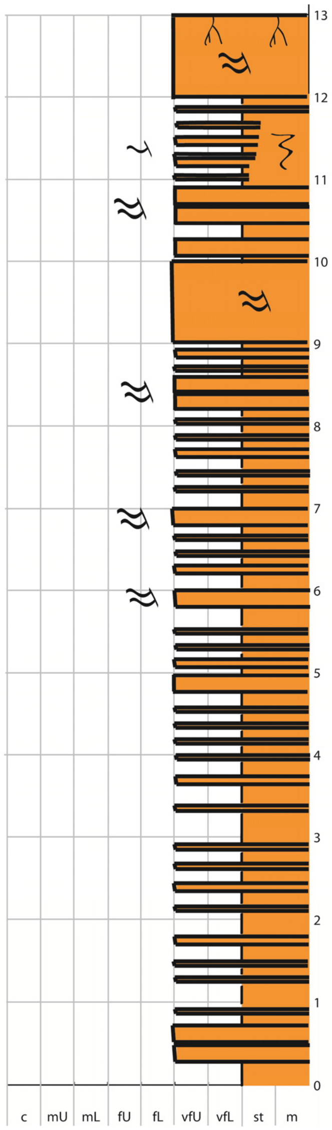
Legend:

- Braided river
- Meandering stream
- Coal
- Overbank
- Swamp
- Ooidal ironstone
- Upper shoreface and foreshore
- Lower shoreface

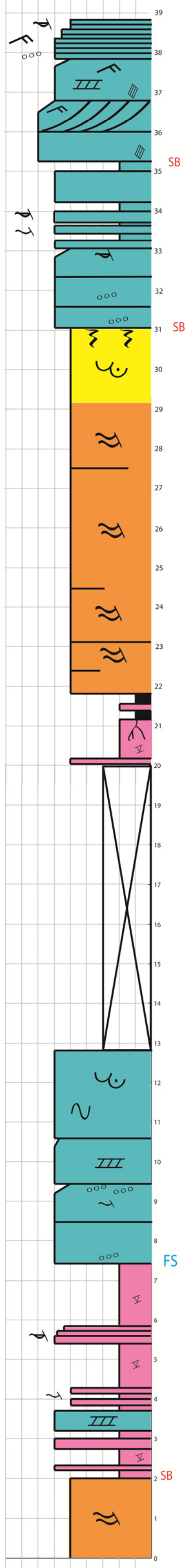
- Tabular cross-bedding
- Wedge cross-bedding
- Festoon cross-bedding
- Planer bedding
- Humocky cross-bedding
- Current ripples
- Wave ripples
- Mud drapes
- Double mud drapes
- Sequence Boundary
- Flooding Surface

- Plant fragments
- Reactivation surface
- Soft sediment deformation
- Bioturbation
- Rooting
- Teredolites
- Rip-up clasts
- Shell lag
- Oolites

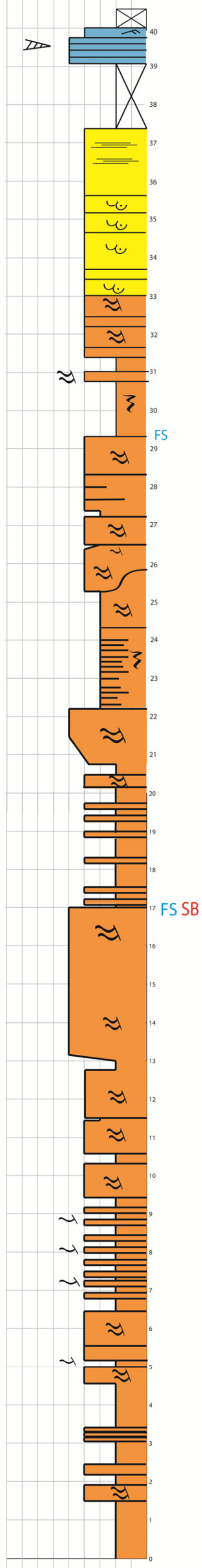
Big Hole Wash



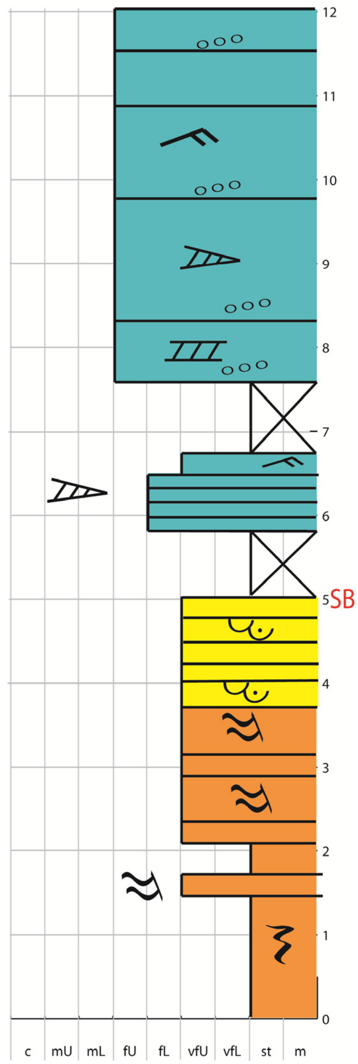
Blaze Canyon



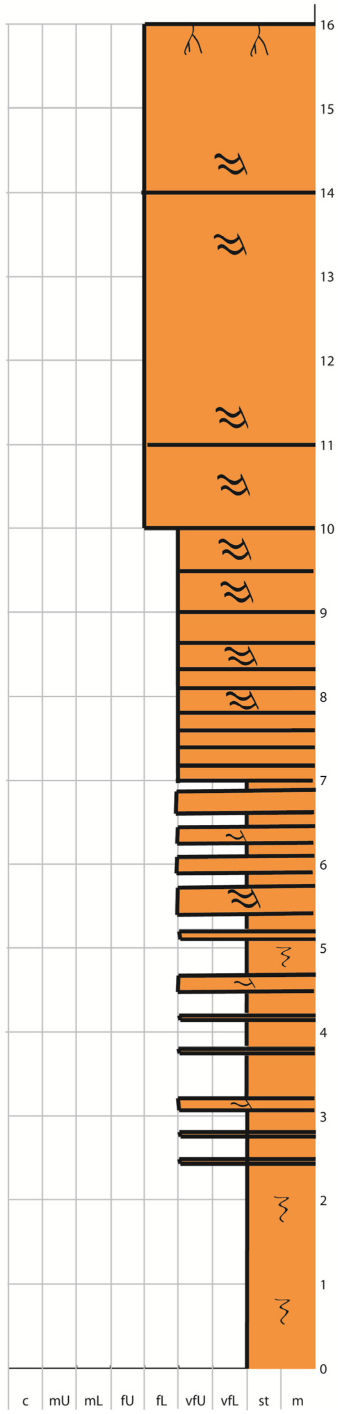
Bull Canyon A



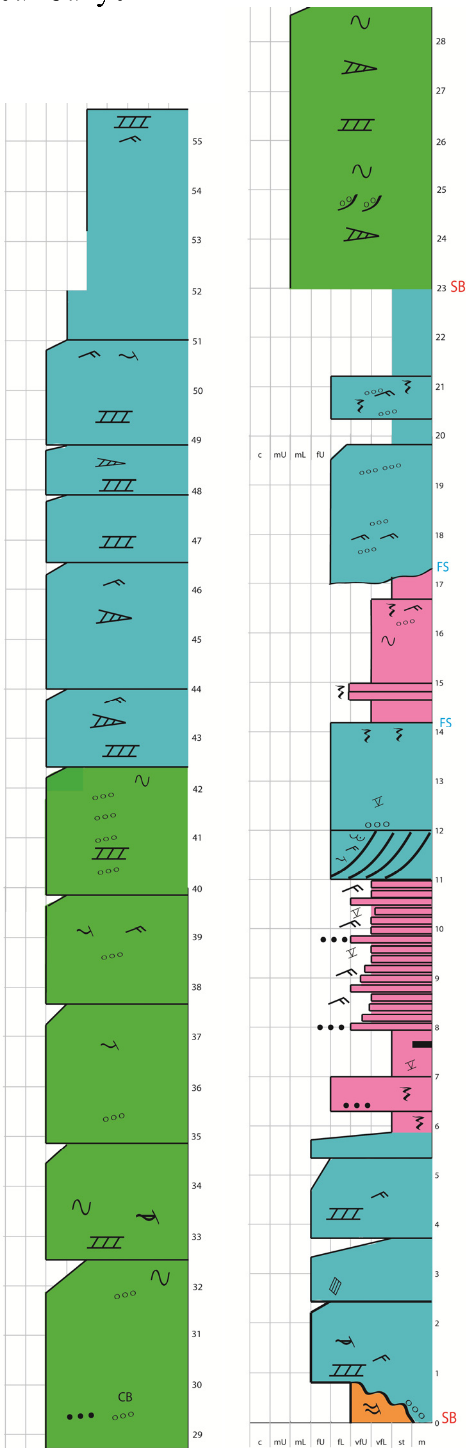
Bull Canyon B



Calf Canyon



Coal Canyon



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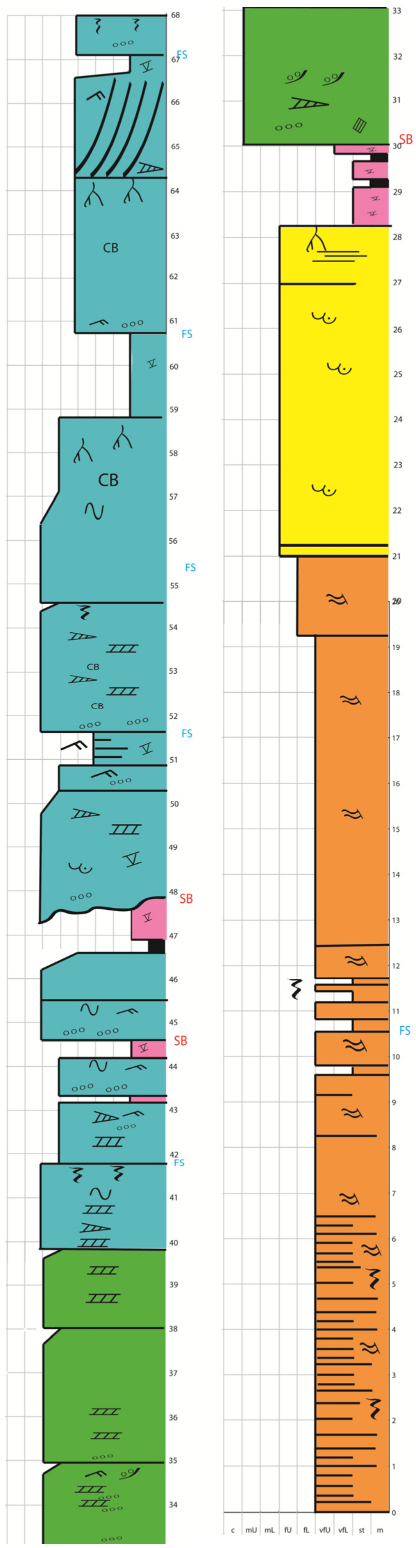
Floy Canyon



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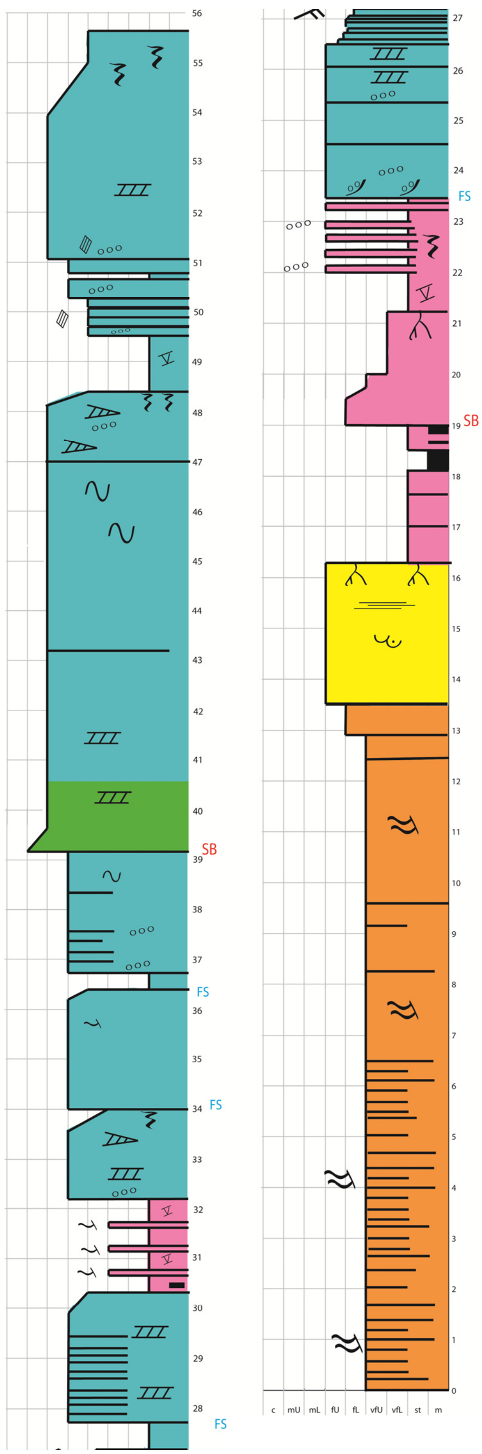
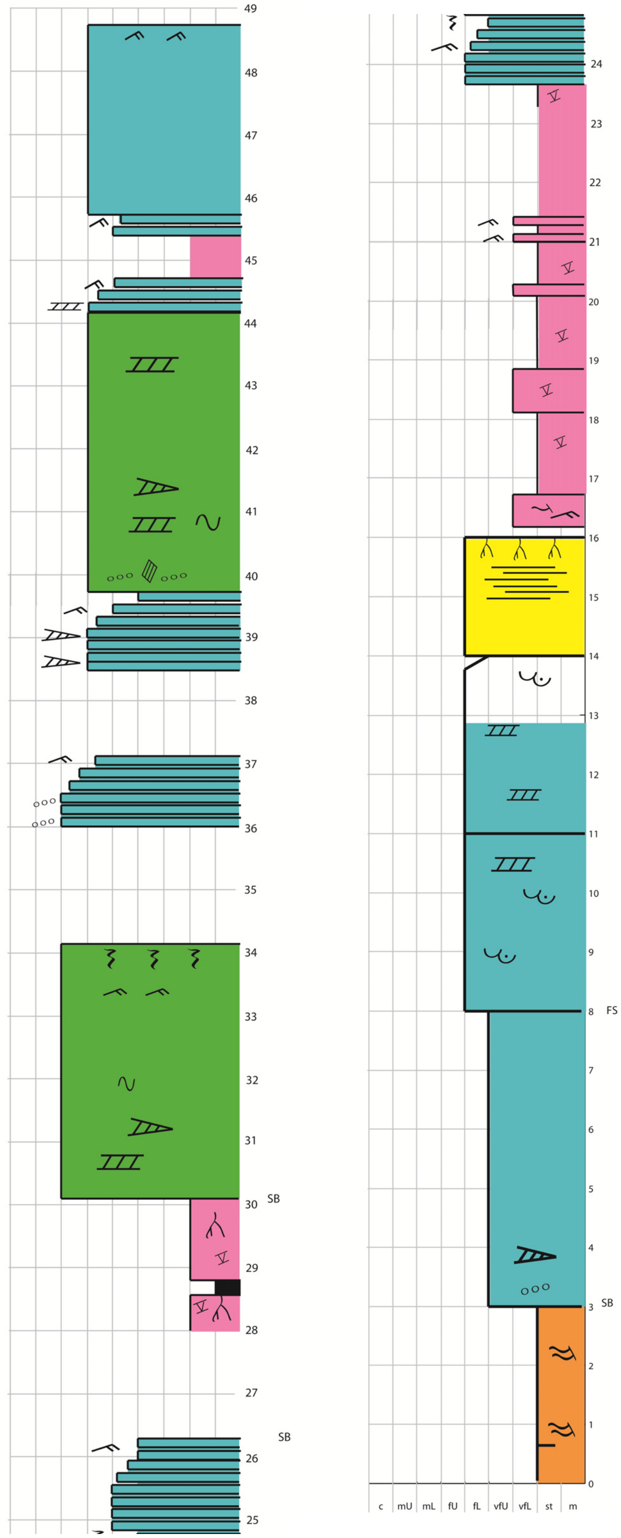


Hatch Mesa



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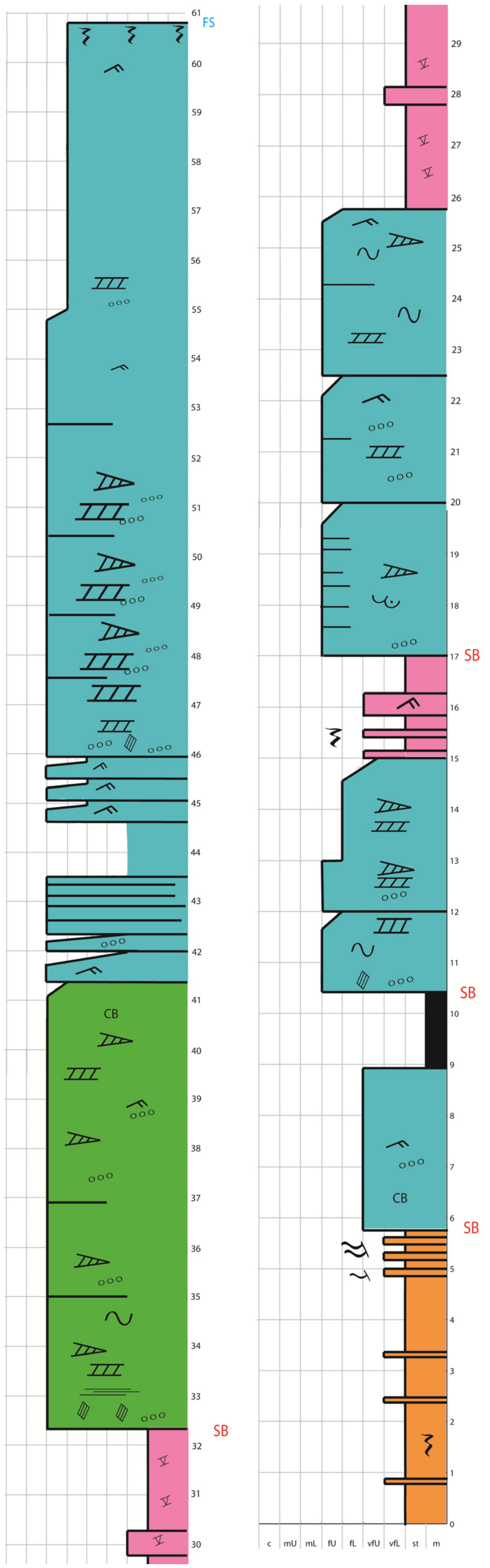
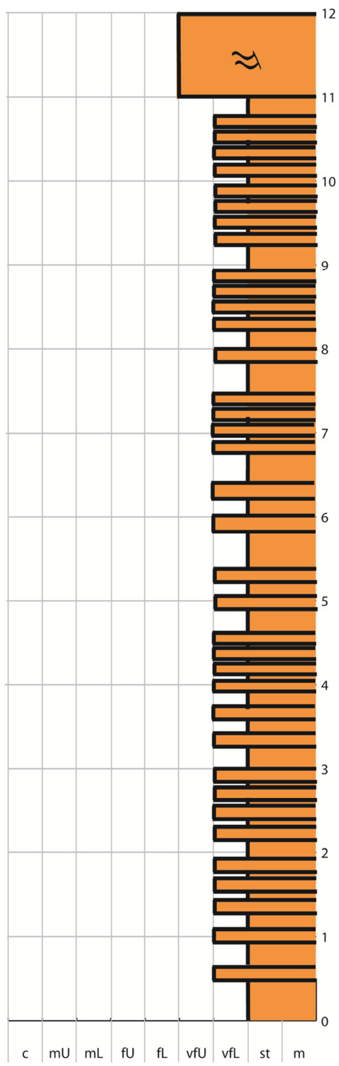
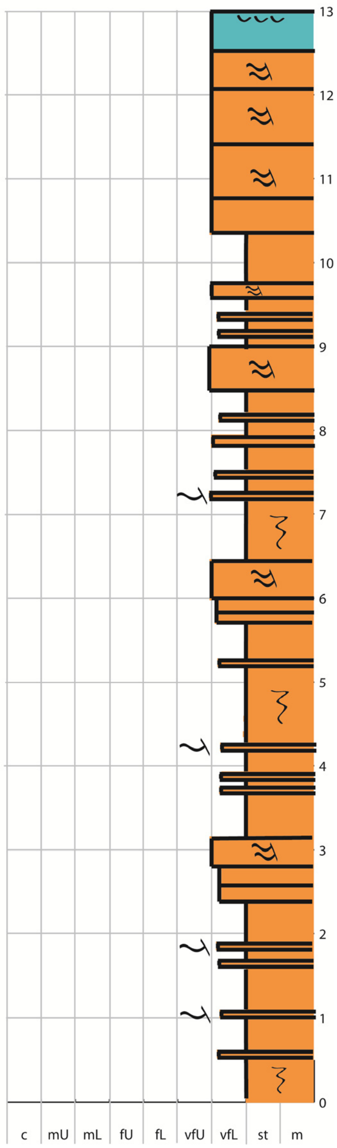
Horse Canyon



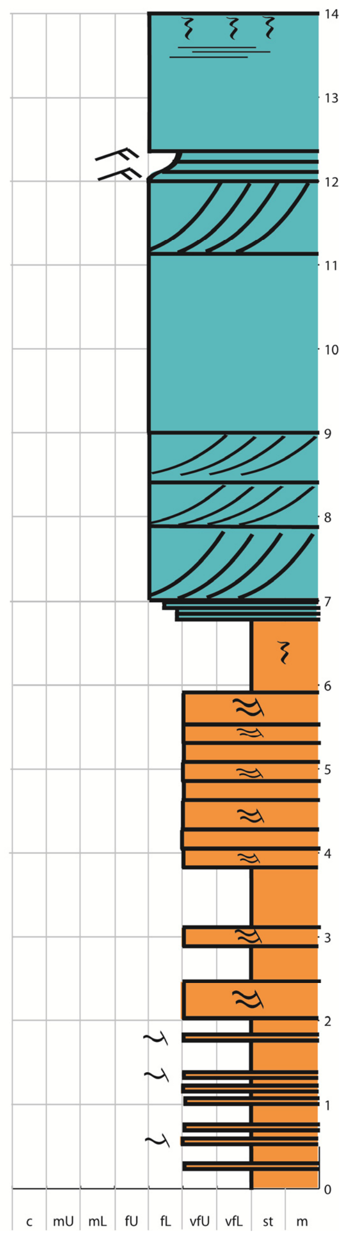
Horse Pastures 1



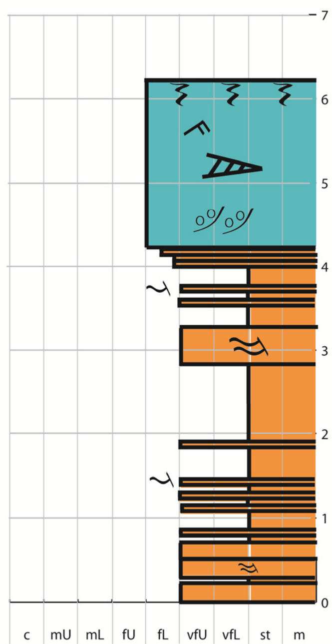
Horse Pastures 2



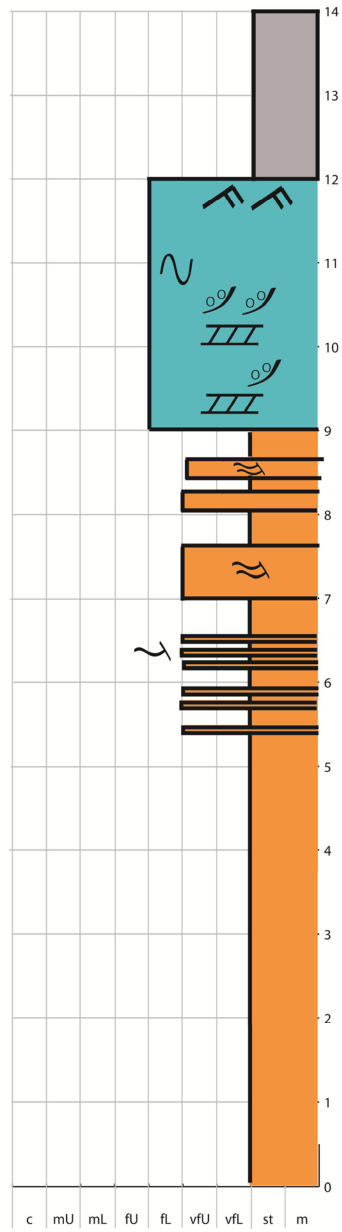
Horse Pastures 3



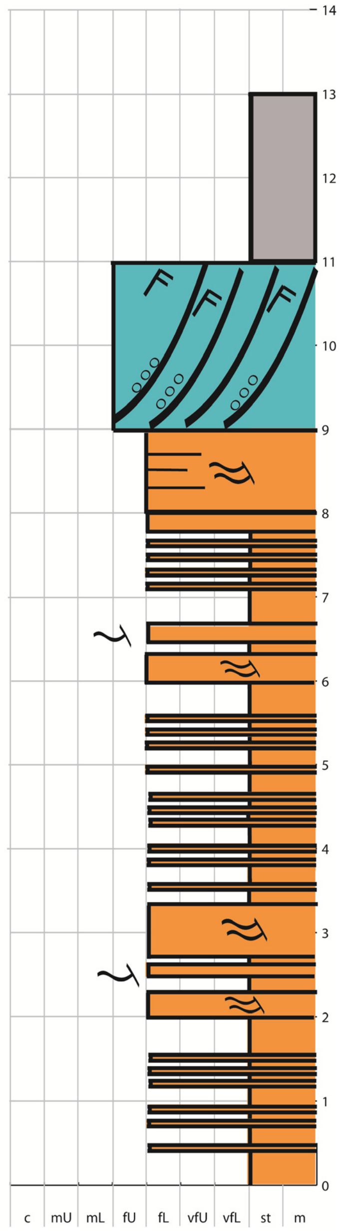
Horse Pastures 4



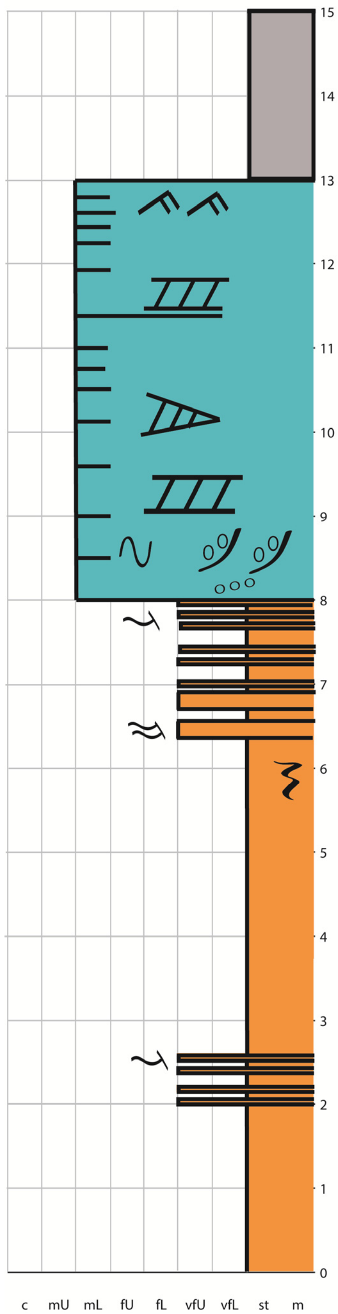
Horse Pastures 5



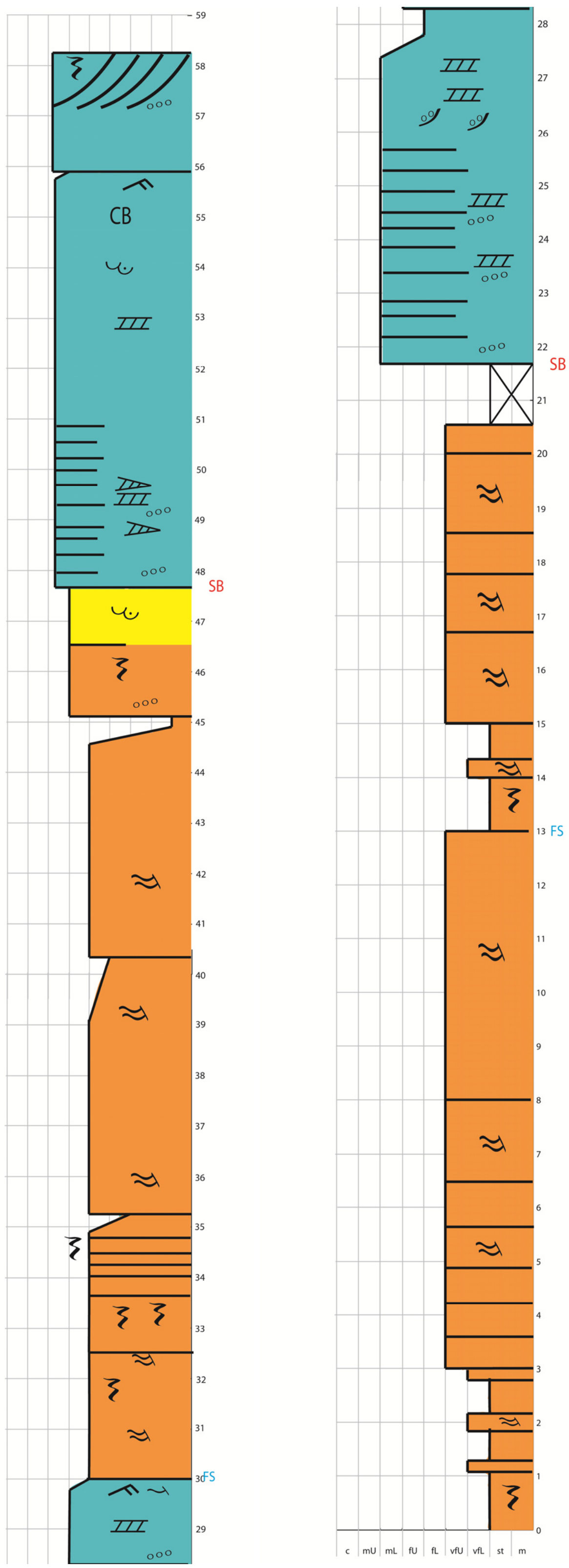
Horse Pastures 6



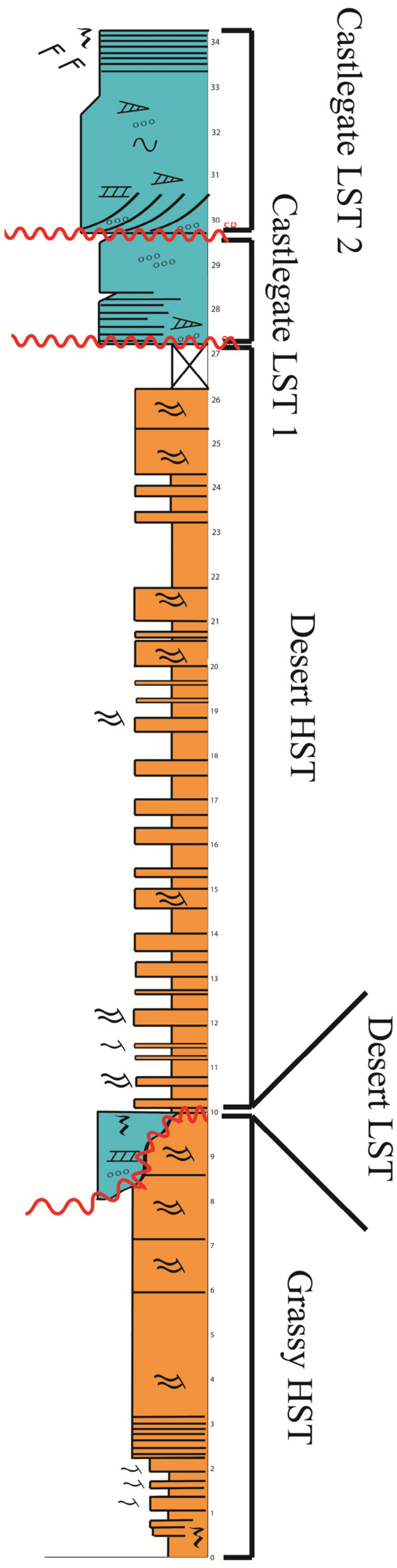
Horse Pastures 7



Jeep Trail

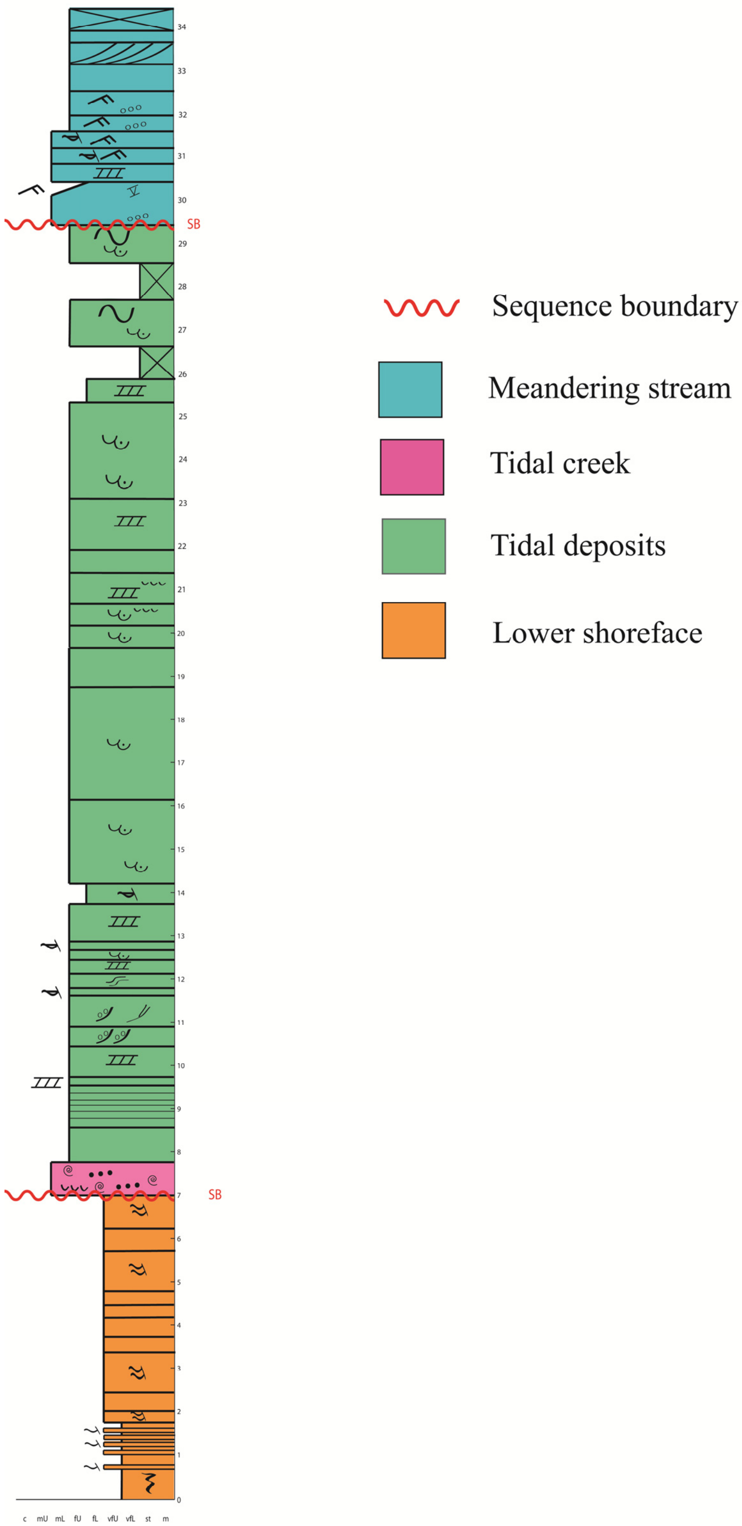


Sagers Canyon

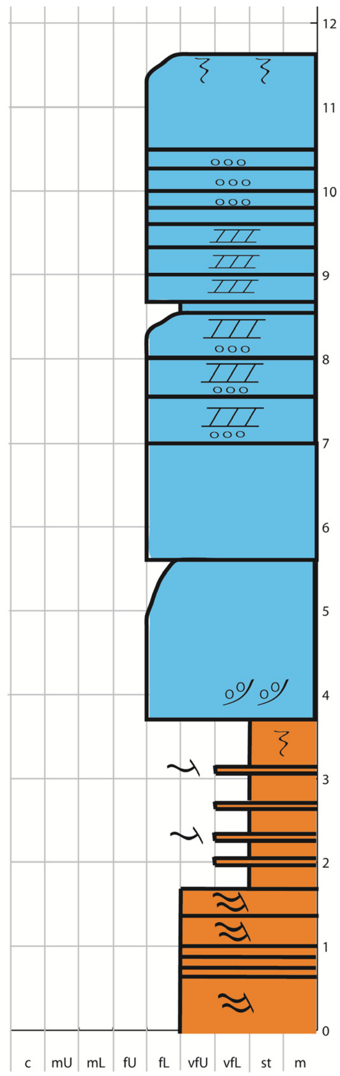


Sagers Canyon East

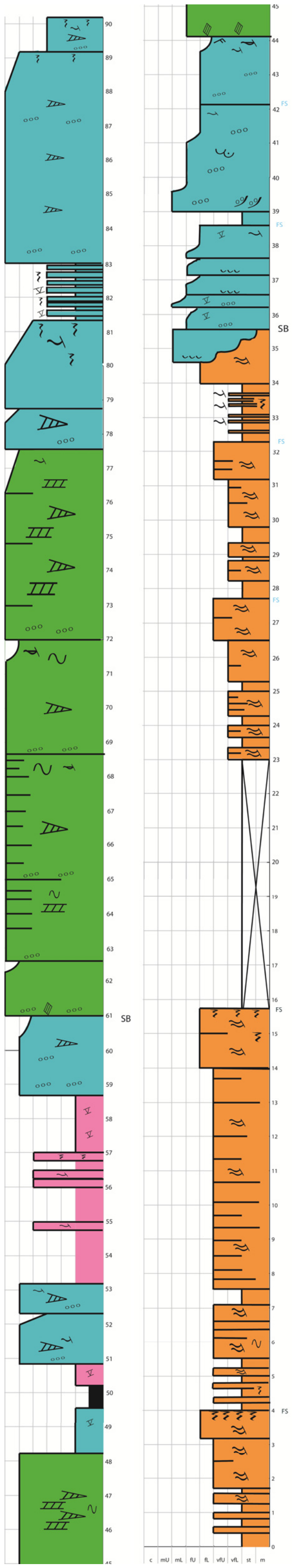
This measured section is not included in the cross section or addressed in the study. Further work in this area is needed to interpret this measured section.



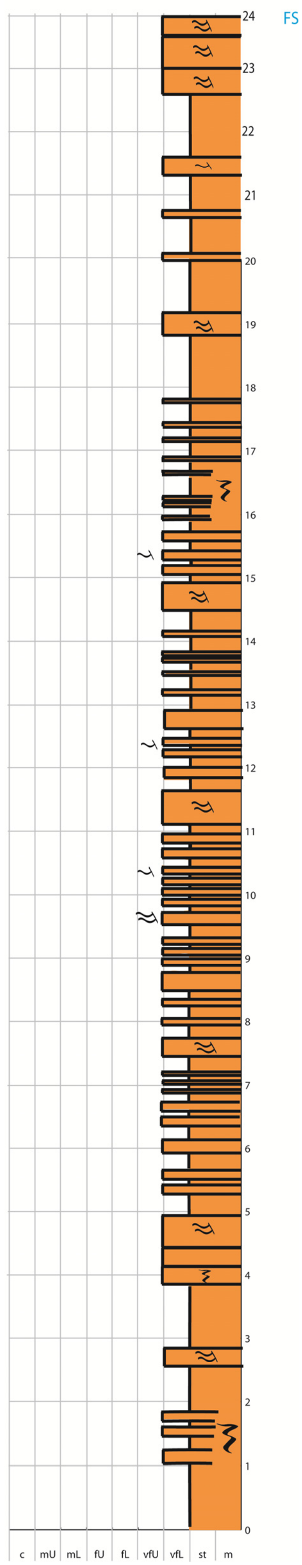
Strychnine Wash



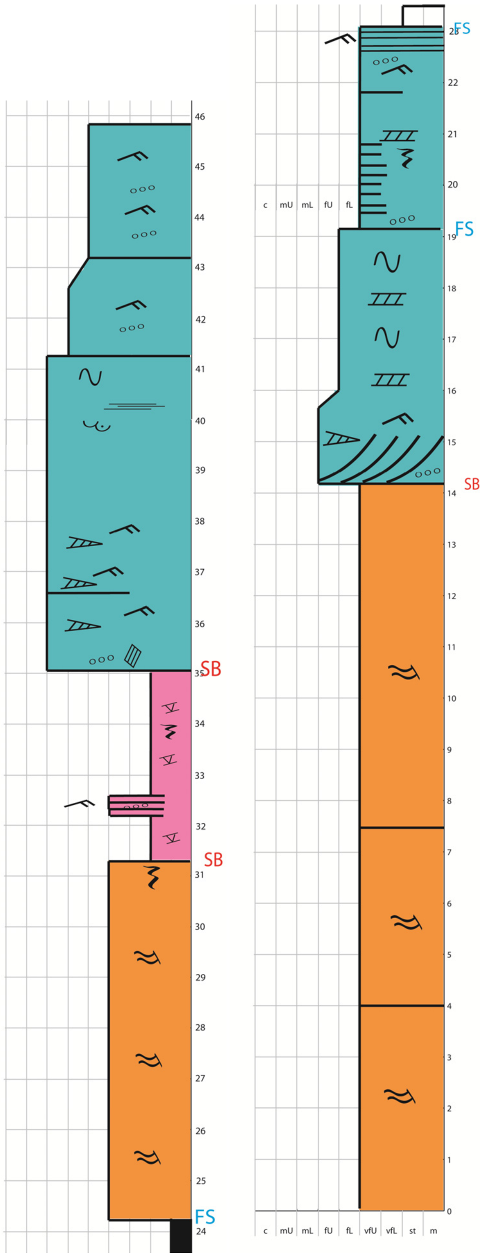
Stubb Canyon



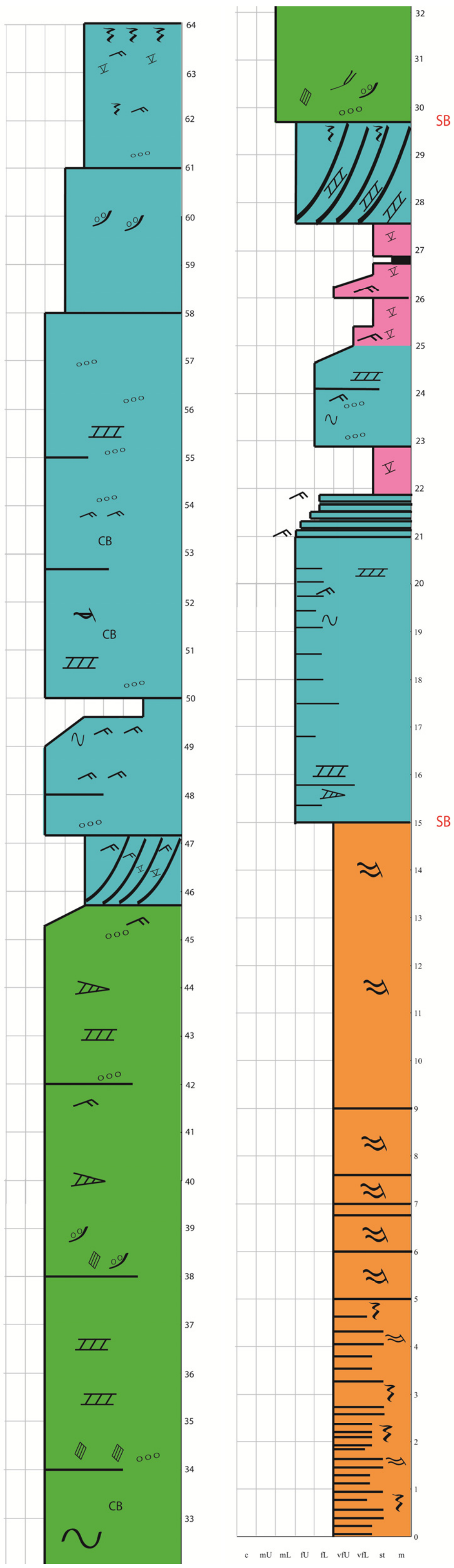
Sulpher Canyon



Thompson Canyon



Tuscher Canyon



West Salt Creek